Modulations of North American and European Weather Variability and Extremes by Interdecadal Variability of the Atmospheric Circulation over the North Atlantic Sector

PATRICK MARTINEAU, HISASHI NAKAMURA, YU KOSAKA, BUNMEI TAGUCHI,^a AND MASATO MORI

Research Center for Advanced Science and Technology, The University of Tokyo, Tokyo, Japan

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ABSTRACT

The dominant mode of wintertime interdecadal covariability between subseasonal surface air temperature (SAT) variability and midtropospheric circulation over the North Atlantic sector is identified through maximum covariance analysis applied to century-long reanalysis data. This mode highlights a tendency for subseasonal temperature variability over Europe and eastern North America to be enhanced during decades when the negative phase of the North Atlantic Oscillation (NAO) prevails. This interdecadal NAO is characterized by a stationary Rossby wave train that originates from the subtropical Atlantic, propagates northward into the subpolar Atlantic, and finally refracts toward Europe and the Middle East. A decadal increase in precipitation in the subtropics under the enhanced supply of heat and moisture from the Gulf Stream and its surroundings appears to act as a source for this wave train. The influence of the interdecadal NAO on subseasonal SAT variability is explained primarily by the modulated efficiency of baroclinic conversion of available potential energy from the background atmospheric flow to subseasonal eddies. The combination of enhanced subseasonal variability and low winter-mean temperature anomalies associated with the negative phase of the interdecadal NAO increases the frequency of cold extremes affecting Europe and the eastern United States.

1. Introduction

Atmospheric variability over the North Atlantic is known to modulate weather and climate over Europe and the east coast of North America (Hurrell et al. 2003). A better grasp of the mechanisms regulating this variability is therefore essential to better predict temperature, precipitation, and storminess affecting these regions. A large fraction of this variability is attributable to the North Atlantic Oscillation (NAO), the dominant mode of atmospheric variability over the North Atlantic sector (Wallace and Gutzler 1981; Ambaum et al. 2001; Barnston and Livezey 1987). The NAO consists of a meridional pressure seesaw over the North Atlantic. By convention, the positive (negative) phase of the NAO represents concomitant

intensification (weakening) of the surface Icelandic low and the Azores high, modulating the climatological pressure gradient. Aloft, the NAO represents a meridional wobbling of the eddy-driven jet stream, with its northward excursion in the positive phase and vice versa (Woollings et al. 2010). NAO variability occurs over a broad range of time scales, from synoptic to decadal scales (Feldstein 2000; Marshall et al. 2001), exerting robust impacts on surface air temperature (SAT) over landmasses surrounding the North Atlantic (Hurrell et al. 2003) and modulating the occurrence of weather extremes (Wettstein and Mearns 2002; Wright et al. 2014; Diao et al. 2015; Luo et al. 2014).

The NAO-related variations on multiple time scales are not necessarily mutually independent. For instance, month-to-month variability of the North Atlantic storm track activity is related to properties of the background westerlies (Wettstein and Wallace 2010; Lau 1988; Rogers 1997). Similarly, Nakamura (1996) showed that year-to-year variability in the seasonal-mean NAO is related to the activity of subseasonal variability over Greenland and Europe. More recently, Woollings et al. (2017) revealed a linkage between interdecadal changes in jet speed and its latitudinal variability on subdecadal time scales: decades with a stronger jet are characterized

⁶ Denotes content that is immediately available upon publication as open access.

^a Current affiliation: Faculty of Sustainable Design, University of Toyama, Toyama, Japan.

Corresponding author: Patrick Martineau, pmartineau@atmos.rcast.u-tokyo.ac.jp

by a steadier jet and vice versa. To explain this relationship, Woollings et al. (2017) found supporting evidence for interdecadal changes in the frequency of blocking events, persistent high pressure anomalies related to wave breaking (Masato et al. 2012), over the Atlantic sector. Since blocking events are linked to the occurrence of weather extremes such as heat waves and cold spells (Black et al. 2004; Dole et al. 2011; Sillmann and Croci-Maspoli 2009; Buehler et al. 2011; Pfahl and Wernli 2012), we speculate that the variability of the North Atlantic circulation on decadal to multidecadal scales can modulate the subseasonal variability of SAT and the occurrence of weather extremes.

The subseasonal and interannual variations of the NAO are driven in part by intrinsic atmospheric processes such as in situ wave breaking (Benedict et al. 2004; Woollings et al. 2008), teleconnections from the North Pacific (Honda et al. 2005a,b), and downward influence from the stratosphere (Baldwin and Dunkerton 2001; Kidston et al. 2015). Prominent external influences include El Niño–Southern Oscillation (Drouard et al. 2015; Jiménez-Esteve and Domeisen 2018) and sea surface temperature (SST) variability over the North Atlantic (Czaja and Frankignoul 1999; Pan 2005; Chen et al. 2020). The interdecadal NAO variability also exhibits a marked linkage with the underlying ocean (Rodwell et al. 1999; Gastineau and Frankignoul 2015; Simpson et al. 2018; Woollings et al. 2015). More specifically, the warm phase of the Atlantic multidecadal oscillation (AMO), characterized by broad warming over the North Atlantic (Deser et al. 2010; Kushnir 1994), is typically associated with the negative phase of NAO and vice versa (Gastineau and Frankignoul 2015; Peings and Magnusdottir 2014). Given this potential influence of interdecadal SST variability on the atmospheric circulation and its modulating effect on the meridional excursions of the jet stream and occurrence of blocking events (Woollings et al. 2017), one may argue that decadal SST variability can affect subseasonal atmospheric variability. In fact, Häkkinen et al. (2011) have found that winters with more frequent blocking over the North Atlantic sector are more likely during the warm phase of AMO. While they argue that it is the change in the blocking frequency that may partly drive multidecadal SST anomalies over the North Atlantic, multidecadal oceanic variability may, in turn, have the potential to modulate blocking frequency as suggested by recent modeling studies (Davini et al. 2015; Peings and Magnusdottir 2014, 2016).

Despite the recent efforts made to deepen our understanding of the processes regulating interdecadal variability over the North Atlantic sector and its influence on subseasonal variability, large uncertainties remain concerning the driving factors of interdecadal atmospheric variability and the dynamical processes by

which it modulates variability on shorter time scales. The present work thus explores, with the use of multidecadal atmospheric reanalysis data, the relationship between the interdecadal variability in the state of atmospheric circulation and subseasonal variability over the North Atlantic sector and seeks a dynamical explanation for the relationship. We first identify the dominant mode of covariability between interdecadal fluctuations of the midtropospheric background state and modulated subseasonal variability through a maximum covariance analysis (see section 2c for details). In the following, fluctuations on time scales between 10 days and a season, from which subweekly variations associated with cyclones and anticyclones migrating along the storm tracks are excluded, are referred to as subseasonal variability. Our focus on this time scale of atmospheric variability is motivated by the need for a deeper understanding of the dynamics of persistent weather anomalies, including blocking events. We also focus on the winter season, when the variability of the North Atlantic atmospheric circulation is largest both on decadal (Simpson et al. 2018) and subseasonal scales (Blackmon 1976; Sheng and Derome 1991; Martineau et al. 2020). Considering its direct socioeconomic impacts, we focus on subseasonal variability of SAT rather than the variance of the midtropospheric circulation as in many other studies (e.g., Blackmon et al. 1984b). Our maximum covariance analysis reveals that interdecadal NAO variability can substantially modulate subseasonal variability. The negative phase of the interdecadal NAO is accompanied by enhanced SAT variability over northeastern North America and Europe. We will then show that interdecadal changes in subseasonal SAT variability are a consequence of amplitude modulations of quasi-stationary eddies. We will also show that the interdecadal NAO circulation anomalies are embedded in a basin-scale wave train originating from the subtropical North Atlantic that appears to be forced in part by precipitation anomalies that are sustained by enhanced turbulent heat fluxes near the Gulf Stream.

2. Data and methods

a. Data

This study primarily uses data from the atmospheric reanalysis of the twentieth century (ERA-20C; Poli et al. 2016) provided by the European Centre for Medium-Range Weather Forecasts (ECMWF) from 1900 to 2010. Unlike conventional reanalyses, ERA-20C only assimilates surface pressure and marine winds. The results based on ERA-20C have been compared (not shown) to those from another reanalysis for the twentieth century, NOAA's Twentieth Century Reanalysis version 2c

(20CRv2c; [Compo et al. 2011](#)), which also limits data assimilation to surface observations. Only the ensemble mean of 20CRv2c is used for the comparison. The ability of these two surface-input reanalyses to capture interdecadal changes in subseasonal surface variability is also briefly evaluated by comparing them to conventional reanalyses: NCEP–NCAR ([Kalnay et al. 1996](#)) and JRA-55 ([Kobayashi et al. 2015](#)), as well as JRA-55C ([Kobayashi et al. 2014](#)), a version of JRA-55 into which satellite-derived data are not assimilated. A thorough description of these datasets is provided in [Fujiwara et al. \(2017\)](#). Whereas only SAT (precisely, temperature 2 m above the surface) is compared among the reanalyses, the in-depth analysis with ERA-20C also uses pressure-level data, including the three-dimensional wind field (u, v, ω), geopotential height (Z), and surface data including surface pressure (p_s), the zonal and meridional components of wind 10 m above the surface (u_{10}, v_{10}), and sensible and latent heat fluxes from the ocean surface. Surface data from ERA-20C are analyzed on a $1.125^\circ \times 1.125^\circ$ horizontal grid, whereas pressure-level data are downsampled to a $3.375^\circ \times 3.375^\circ$ horizontal grid and use 19 equally spaced pressure levels from 1000 to 100 hPa for the analysis of atmospheric energetics described below.

In addition, monthly SST fields with a $1^\circ \times 1^\circ$ resolution from the HadISST dataset produced by the Met Office Hadley Centre ([Rayner et al. 2003](#)) are used. The analysis is also carried out with the Extended Reconstructed Sea Surface Temperature (ERSST) v5 ([Huang et al. 2017](#)) dataset without showing any discernible sensitivity (not shown).

b. Index of subseasonal surface temperature variability and its interdecadal fluctuations

The subseasonal SAT variability is evaluated by applying a 10–60-day Butterworth bandpass filter locally for each winter season from November to March, after which only the months of December–February (DJF) are retained to eliminate artifacts arising from the filter edges. Whereas the 10-day lower bound is often used to separate subseasonal variability from high-frequency variability ([Kushnir and Esbensen 1986](#); [Blackmon et al. 1984a](#); [Nakamura 1996](#)), there is no established standard for its upper bound. Here we use an upper limit of 60 days, similar to [Kushnir and Wallace \(1989\)](#), to ensure a clear separation from seasonal-scale variability, which is important for investigating interactions between the wintertime basic-state circulation and subseasonal variability. The separation of the subseasonal eddies from the high-frequency eddies is justified by the differences in wave–mean flow interactions between the two time scales ([Hoskins et al. 1983](#); [Sheng and Derome 1991](#); [Lau and Nath 1991](#)) and the need to better understand the dynamics of persistent weather extremes. Similar results are obtained when using a 10–90-day filter (not shown).

To characterize how subseasonal SAT variability is modulated from one year to another, we locally define a measure of subseasonal variability (SSV) as the standard deviation of the subseasonal SAT component for every winter season. Interdecadal modulations in SSV are represented by smoothing the SSV time series with a 10-yr Butterworth low-pass filter. The resultant index is hereafter referred to as interdecadal modulations of subseasonal variability (SSV_{ID}). Both SSV and SSV_{ID} are illustrated in [Fig. 1](#) at several locations around the North Atlantic sector. The time series are representative of an average over a 10° longitude $\times$ 10° latitude box centered at each reference location to reduce interdataset bias arising from the differences in horizontal grids among reanalysis datasets. The locations that tend to be affected strongly by interdecadal changes of SSVID around the North Atlantic sector ([section 3b](#)) are chosen. SSV varies substantially from one year to another and, at some locations, almost doubles between its maximum and minimum. The interdecadal variability isolated with SSV_{ID} is not as large as the full interannual variability but nonetheless shows impressive extents in some locations such as over Quebec and Denmark. The SSV and SSV_{ID} indices are similar between surface-input (ERA-20C and 20CRv2c) and conventional reanalyses (NCEP–NCAR and JRA-55), despite some regional mismatches (especially over Denmark for ERA-20C). This suggests that subseasonal variability and its year-to-year modulations are reasonably well constrained by the assimilation of surface data alone. We note that ERA-20C and 20CRv2c diverge over Quebec and Denmark before the 1970s. This appears to be largely due to a time-mean bias since the two reanalyses overall exhibit similar interdecadal variability. We have repeated the same analysis of SSV_{ID} with 20CRv2c and obtained similar results (not shown), indicating that this bias does not adversely affect our analysis of interdecadal variability. We select ERA-20C for our main analysis primarily because it has a better representation of the dynamical variability in the upper troposphere ([Gerber and Martineau 2018](#)), which is necessary for evaluating the energetics of atmospheric eddies.

c. Maximum covariance analysis

Leading modes of interdecadal covariability between the winter-mean atmospheric circulation and SSV are identified with maximum covariance analysis ([Bjornsson and Venegas 1997](#); [Bretherton et al. 1992](#)) by applying singular value decomposition analysis (SVD) to the covariance matrix between the 10-yr low-pass-filtered time series of DJF-mean geopotential height at 500 hPa ($Z500_{ID}$) and the SSV_{ID} . Through the SVD analysis, the component of subseasonal variability (SSV_{ID}) that is modulated by changes in large-scale atmospheric circulation ($Z500_{ID}$) is identified. The covariance matrix is

FIG. 1. The evolution of the subseasonal variability (SSV) of SAT is shown for several reference locations (rows) and various reanalysis datasets (color-coded according to the legend). The reference coordinates are indicated in the upper-left corner of each panel. The SSV time series are shown with thin lines and their decadal fluctuation (SSV_{ID}), obtained by applying a 10-yr low-pass filter to SSV, is shown with thick lines. All values have units of K. Note that JRA-55 and JRA-55C are almost indistinguishable.

constructed for the period from 1905 to 2005 and the domain 100°W – 20°E , 5° – 65°N over the midlatitude North Atlantic. The results are insensitive to the choice of the domain. In fact, the dominant SVD mode over the North Atlantic can be retrieved from a similar SVD analysis carried out for the whole Northern Hemisphere, although the fraction of the total squared covariance explained is reduced substantially. The SVD analysis applied to the correlation matrix in place of the covariance matrix yields essentially the same results. The use of Z500 to characterize the basic-state circulation is motivated by the fact that subseasonal variability is produced by eddies that have deep vertical structures (Blackmon et al. 1979; Taguchi and Asai 1987) and derive their energy from the mean flow within the whole troposphere (Sheng and Derome 1991; Cai et al. 2007). A similar SVD analysis with sea level pressure in place of Z500 yields consistent results (not shown).

d. Transfers of energy between the mean flow and subseasonal variability

Interdecadal changes in the energy of subseasonal eddy anomalies can be quantified by evaluating the

atmospheric energy budget. It was introduced by Lorenz (1955) with a decomposition of the atmospheric circulation into its zonal mean and deviations from it and was later reformulated by Oort (1964) with a partitioning of energy associated with the time-mean flow and anomalies. Our study uses the latter to investigate sources of eddy energy over the subseasonal frequency band. We define the subseasonal component of variability, denoted here with primes, in the same manner as we defined the SSV index (10–60-day bandpass filter). The slowly evolving basic state, denoted with overbars, is defined as the DJF mean in each year. For the purpose of this study, which focuses on interdecadal variability, a 10-yr low-pass filter is further applied on the winter-mean fields.

The eddy kinetic energy (EKE) of subseasonal anomalies is defined as

$$EKE = \frac{u'^2 + v'^2}{2}. \quad (1)$$

The barotropic conversion of kinetic energy (CK) from the background state to subseasonal eddies is expressed as

$$CK = \frac{v'^2 - u'^2}{2} \left(\frac{\partial \bar{u}}{\partial x} - \frac{\partial \bar{v}}{\partial y} \right) - u'v' \left(\frac{\partial \bar{u}}{\partial y} + \frac{\partial \bar{v}}{\partial x} \right). \quad (2)$$

The first term is positive for zonally elongated eddies ($v'^2 < u'^2$) embedded in a jet exit region ($\partial u / \partial x < 0$) (Simmons et al. 1983; Mak and Cai 1989; Kosaka and Nakamura 2006). The eddy available potential energy (EAPE) associated with subseasonal thermal anomalies is defined as

$$EAPE = \gamma^{-1} \frac{T'^2}{2}. \quad (3)$$

The stability factor γ is written as

$$\gamma = \frac{p}{R} \left(\frac{\widehat{R\bar{T}}}{C_p p} - \frac{\widehat{\partial \bar{T}}}{\partial p} \right), \quad (4)$$

where R is the gas constant of dry air ($287 \text{ J K}^{-1} \text{ kg}^{-1}$), C_p is the specific heat of air at constant pressure ($1004 \text{ J K}^{-1} \text{ kg}^{-1}$), and the hat operator denotes horizontal averaging over the Northern Hemisphere. The conversion of available potential energy (CP) from the background flow to subseasonal eddies may be expressed (Kosaka and Nakamura 2006; Tanaka et al. 2016) as

$$CP = -\gamma^{-1} \left(u' T' \frac{\partial \bar{T}}{\partial x} + v' T' \frac{\partial \bar{T}}{\partial y} \right), \quad (5)$$

which is positive when eddy heat fluxes are directed down the gradient of the basic-state temperature. Transfer from EAPE to EKE is evaluated with

$$CPK = -\frac{R\omega' T'}{p}. \quad (6)$$

Feedbacks from high-frequency eddies, which are extracted with a 10-day high-pass Butterworth filter and denoted with double primes, are also evaluated with the method of Tanaka et al. (2016) with

$$CP_{\text{HF}} = -\gamma^{-1} T' \left[\frac{\partial(u'' T'')}{\partial x} + \frac{\partial(v'' T'')}{\partial y} \right], \quad (7)$$

which quantifies how high-frequency eddy heat flux convergence reinforces subseasonal thermal anomalies, and

$$CK_{\text{HF}} = -u' \left[\frac{\partial(u'' u'')}{\partial x} + \frac{\partial(u'' v'')}{\partial y} \right] - v' \left[\frac{\partial(u'' v'')}{\partial x} + \frac{\partial(v'' v'')}{\partial y} \right], \quad (8)$$

which quantifies how high-frequency eddy momentum fluxes reinforce subseasonal wind anomalies.

Finally, diabatic sources or sinks of EAPE are evaluated with

$$CQ = \gamma^{-1} \frac{Q' T'}{C_p}, \quad (9)$$

where the subseasonal component of diabatic heating Q' has been obtained as the residual of the thermodynamic equation since parameterized heat tendencies are not provided for ERA-20C. In practice, all the conversion terms in (2), (5), (7), and (8) were evaluated with the formulas for spherical coordinates and subjected to a 10-yr low-pass filter.

Attempting to explain changes in eddy energy with the use of energy conversion terms can lead to some ambiguity since both the energy terms (EKE and EAPE) and conversion terms (CP and CK) depend on the squared magnitude of eddy quantities. To circumvent this ambiguity, we evaluate the efficiency of energy conversion as the ratio of the conversion with respect to the total eddy energy (EAPE + EKE). For instance, the efficiency of CP is obtained with $CP^{\text{eff}} = CP / (EAPE + EKE)$ after evaluating the hemispheric and vertical integrals of CP, EAPE, and EKE. The efficiencies are computed in a similar manner for CK^{eff} , $CK_{\text{HF}}^{\text{eff}}$, $CP_{\text{HF}}^{\text{eff}}$, and CQ^{eff} . The unit of the efficiencies is per day (day^{-1}). The inverse of the efficiency thus offers a measure of the time scale needed to fully replenish the energy of subseasonal eddies (Kosaka and Nakamura 2006).

e. Statistical significance

Statistical significance for all regressions is evaluated with a t test for the population correlation coefficient r . The test statistic takes the form $t = r\sqrt{N}/\sqrt{1-r^2}$ (von Storch and Zwiers 1999), where N is the number of degrees of freedom. Here we estimate the effective number of degrees of freedom taking into account serial correlation as

$$N_{\text{eff}} = n / \left[1 + 2 \sum_{\tau=1}^{\tau_{\text{max}}} (1 - \tau/n) r_X(\tau) r_Y(\tau) \right], \quad (10)$$

where $r_X(\tau)$ and $r_Y(\tau)$ are the autocorrelations of the two time series considered at lag τ (Bretherton et al. 1999; Metz 1991).

f. AMO index

For our investigation, the AMO index is obtained through the same method as described in detail by Ting et al. (2009), except using winter-mean SST only. Since the AMO index does show some sensitivity to the season

FIG. 2. Decadal standard deviation (shadings) and climatology (contours) of DJF mean statistics. The variables shown are the zonal wind at 700 hPa (U700), the eddy kinetic energy of subseasonal eddies at 700 hPa (EKE700), the eddy available potential energy of subseasonal eddies at 700 hPa (EAPE700), and the subseasonal standard deviation of SAT (SATVAR; also referred to as SSV). Stippling indicates regions where the interdecadal variability is significantly different from white noise at the 95% confidence level. Contours of standard deviation and climatology are in steps of 0.4 and 2.5 m s^{-1} for U700, 1 and 10 J kg^{-1} for EKE700, 7.5 and 30 J kg^{-1} for EAPE700, and 1 and 15 K for SATVAR.

used to compute it (Yamamoto and Palter 2016), we focus on the winter index to more accurately capture interactions with the winter-mean atmospheric circulation. This is also motivated by the fact that the connection between SST variability and atmospheric circulation variability over the North Atlantic sector is seasonally dependent, with a maximum in late winter (Simpson et al. 2018). Nevertheless, interdecadal modulations of subseasonal variability are found to maximize in midwinter (not shown), and thus we focus on DJF for our analysis. Specifically, the first step consists of a grid-by-grid removal of the global warming SST signal by linearly regressing local SST anomalies onto SST averaged from 60°S to 60°N. The AMO index is then obtained as the area-averaged residual of the linear regression over the North Atlantic sector (0°–60°N, 80°W–0°). A low-pass filter with a 10-yr cutoff is subsequently applied and followed by normalization.

3. Results

a. An overview of interdecadal variability

The interdecadal variability of the lower-tropospheric (700 hPa) circulation and activity of subseasonal eddies, with periods ranging from 10 to 60 days, is overviewed as wintertime statistics for DJF (Fig. 2), when the eddy-driven jet (U700) over the North Atlantic and its interdecadal variability are stronger than in any other seasons. The climatology and interdecadal variability of EKE are both

maximized over the North Atlantic sector, where the exit region of the eddy-driven jet is located, in agreement with classical analyses of subseasonal variability (Blackmon et al. 1977, 1984a; Holopainen 1984). Meanwhile, the EAPE of subseasonal eddies is the largest over high-latitude landmasses but weaker over the ocean, which is indicative of damping of lower-tropospheric thermal variability at the ocean–atmosphere interface. The interdecadal variability of EAPE700 is also largest over the same regions. Likewise, SAT variability (SATVAR) is climatologically larger over land than over the ocean, and its decadal modulations are also larger over the landmasses.

By generating 10 000 synthetic Gaussian white noise time series with the same variance as the original time series, we have verified that interdecadal variations in U700, EKE, EAPE, and SATVAR are unlikely to be explained by white noise over some regions of the North Atlantic sector and surrounding landmasses. This result indicates that their variability is unlikely to be purely internally driven and likely to be under an external influence such as ocean variability (Simpson et al. 2018). Similar results are obtained when assuming a red noise process instead of white noise since 1-yr lag autocorrelation of each of those atmospheric variables investigated is typically quite small (not shown).

b. Maximum covariance analysis

The dominant mode of covariability (SVD1) between SSV_{ID} and Z500_{ID} (Fig. 3) explains 87% of the squared

FIG. 3. The leading mode of the SVD analysis (SVD1) applied between SSV_{ID} and 10-yr low-pass-filtered 500-hPa geopotential height ($Z500_{ID}$). Heterogeneous regressions are shown for (a) SSV_{ID} and (b) $Z500_{ID}$ (featuring the negative phase of interdecadal NAO). Statistically significant regions at the 95% level are contoured with magenta lines. The fraction of the variance of the original fields explained by SVD1 (i.e., square of heterogeneous correlations) is contoured in green with intervals of 20%. The expansion coefficients corresponding to the SSV_{ID} pattern ($SVD1_{SSVID}$; red) and the $Z500_{ID}$ pattern ($SVD1_{Z500ID}$; blue) are shown in (c). The AMO index and the sign-reversed NAO_{ID} index (note that NAO_{ID} is often indistinguishable from $SVD1_{Z500ID}$ or $SVD1_{SSVID}$) are also illustrated according to the color legend.

covariance between the two variables over the North Atlantic sector. In contrast, the second SVD mode (not shown) explains only 7% of the squared covariance, and its impact on SSV_{ID} is weak and of smaller spatial scales. SVD1 is characterized by an enhancement of SSV_{ID} over northeastern North America, Europe, and northern Africa (Fig. 3a), in association with dipolar $Z500_{ID}$ anomalies over the North Atlantic with its anticyclonic center over Greenland and cyclonic anomalies stretching zonally over the midlatitude North Atlantic (Fig. 3b). The two expansion coefficients, $SVD1_{SSVID}$ and $SVD1_{Z500ID}$, obtained by projecting the original variables onto the loading patterns, are highly correlated (see Fig. 3c and Table 1). SVD1 explains a substantial fraction of the total variability of SSV_{ID} with up

to 40% of the variance explained over northeastern North America and parts of Europe. SVD1 also explains a substantial fraction of $Z500_{ID}$ variability over Greenland ($\sim 40\%$) and Europe ($\sim 40\%$) but even more so in the eastern subtropical Atlantic ($\sim 60\%$), which suggests that SVD1 is strongly tied to atmospheric variability over the subtropical Atlantic sector (Fig. 3b).

The dipolar $Z500_{ID}$ anomalies shown in Fig. 3b are similar to the negative phase of the NAO (NAO $-$; Woollings et al. 2010), except that the southern negative pole stretches more zonally and shows two distinct maxima: one over Europe and the other over the western subtropical Atlantic. This anomaly pattern is actually similar to the dominant EOF of $Z500_{ID}$ (NAO $_{ID}$; not shown) and the time series of the sign-reversed

TABLE 1. Correlation coefficients between the expansion coefficients obtained from the SVD analysis ($SVD1_{SSV_{ID}}$ and $SVD1_{Z500_{ID}}$), the sign-reversed interdecadal North Atlantic Oscillation index ($-NAO_{ID}$), the principal component time series of the leading EOF of SSV_{ID} , and the AMO index. Correlations that are significant at the 95% level are listed in boldface.

	$SVD1_{Z500_{ID}}$	$-NAO_{ID}$	EOF1 (SSV_{ID})	AMO
$SVD1_{SSV_{ID}}$	0.84	0.78	0.95	0.43
$SVD1_{Z500_{ID}}$		0.94	0.79	0.48
$-NAO_{ID}$			0.71	0.49
EOF1 (SSV_{ID})				0.44

NAO_{ID} (Fig. 3c) is strongly correlated to $SVD1_{Z500}$ (Table 1), indicating that SVD1 essentially expresses the interdecadal component of NAO variability. Annular variability of the NH circulation, referred to as the Arctic Oscillation (AO) or the northern annular mode (NAM), is also highly correlated with this mode, given its strong connection with the NAO (Deser 2000; Ambaum et al. 2001). The difference between the spatial $Z500_{ID}$ structure of SVD1 and the structure of NAO defined from monthly and subseasonal data (Woollings et al. 2010; Martineau et al. 2020) may hint to potential differences in the dynamical mechanisms driving the variability of the atmospheric circulation over the North Atlantic sector between these time scales. Due to the high degree of similarity, we will hereafter regard SVD1's $Z500_{ID}$ signature as interdecadal NAO variability, keeping in mind that a positive $SVD1_{Z500_{ID}}$ corresponds to the negative phase of NAO_{ID} . Likewise, the SSV_{ID} pattern is similar to the leading EOF of SSV_{ID} field [EOF1(SSV_{ID})], and the $SVD1_{SSV_{ID}}$ and EOF1(SSV_{ID}) time series are well correlated (Table 1). SVD1 thus uncovers a robust statistical connection between the respective dominant modes of variability in SSV_{ID} and $Z500_{ID}$.

Typical time scales of the SVD1 variability are examined by evaluating its power spectral characteristics. The interannual variability in SSV and Z500 associated with SVD1 is obtained by projecting unfiltered SSV and Z500 anomalies onto their respective loading patterns of SVD1. The resulting time series, which show substantial year-to-year fluctuations, are used to assess whether the variability of $SVD1_{SSV_{ID}}$ and $SVD1_{Z500_{ID}}$ is explained by the interdecadal sampling of white noise. To this end, the power spectra of 10 000 synthetic Gaussian white noise time series with the same variance as the interannually varying time series have been generated to construct the 95% and 99% confidence intervals. As shown in Fig. 4, several peaks of the $SVD1_{SSV_{ID}}$ and $SVD1_{Z500_{ID}}$ spectra rise above these confidence levels. One of them is found for $SVD1_{Z500_{ID}}$ between 10 and 15 years, and two peaks are found for $SVD1_{SSV_{ID}}$ between

10 and 15 years and at about 20 years. These significant spectral peaks suggest that an external forcing may influence SVD1 on decadal to multidecadal time scales.

Among the external forcings that are known to influence North Atlantic atmospheric variability, the AMO is one of the main contenders (Deser et al. 2010; Gastineau and Frankignoul 2015; Peings and Magnusdottir 2014; Simpson et al. 2018). Whereas the AMO index (Fig. 3c) varies mostly on longer multidecadal time scales, the SVD1 time series exhibits much stronger variability at periods of 10–30 years (Figs. 3 and 4). As a result, the correlation between the AMO and the SVD1 time series is rather modest (Table 1). The possibility of an influence from the ocean will be discussed further in the next section.

c. Atmospheric and oceanic variability and interactions on decadal to multidecadal scales

This section investigates the NAO-like atmospheric anomalies associated with SVD1 in more detail. To this end, upper-tropospheric vorticity anomalies are used to identify possible teleconnections, or wave trains, involved with the NAO-. The top panel of Fig. 5 shows the anomalous 10-yr low-pass filtered winter-mean relative vorticity regressed onto $SVD1_{Z500_{ID}}$ and associated fluxes of stationary Rossby wave activity (Takaya and Nakamura 2001) evaluated from the regressed height anomalies at 300 hPa. The figure depicts a wave train whose wave activity propagates poleward over the North Atlantic before being refracted toward Europe and the Middle East. The wave-activity flux is divergent over the subtropical western North Atlantic (around 35°N), suggesting the origin of this wave train. The circulation anomalies illustrated with the relative vorticity in Fig. 5 are not confined to the upper troposphere but extend down to the surface with an almost equivalent barotropic structure (not shown). In the extratropics, one may notice that the positive (cyclonic) vorticity anomaly is organized in a zonally elongated band with two distinct centers, similar to the $Z500_{ID}$ pattern shown in Fig. 3b. The wave-activity flux indicates that both centers are part of the wave train. While the western center is associated with the poleward-propagating branch, the eastern center is associated with the south-eastward propagating branch. Because of the long time scale involved, this wave train can be considered as a stationary wave train that is likely externally forced.

To better understand the origin of the stationary wave train, the Rossby wave source (RWS) is evaluated (Sardeshmukh and Hoskins 1988). It is expressed as $-\nabla \cdot [\mathbf{u}_\chi(\zeta + f)]$, where $\mathbf{u}_\chi$ is the divergent wind component, ζ the relative vorticity, and f the Coriolis parameter. RWS has been computed with monthly-mean data and then regressed onto $SVD1_{Z500_{ID}}$. The RWS is responsible

FIG. 4. (top) Time series with full interannual variability (thin lines) are obtained by projecting the unfiltered Z500 and SSV fields onto their respective SVD1 loading patterns. They are shown together with the expansion coefficients (thick lines) shown previously in Fig. 3. (bottom) The power spectral density of the interannual time series is shown. Dotted and dashed lines indicate the 95% and 99% confidence levels for the power spectrum of Gaussian white noise with the same standard deviation as the interannually varying time series. The power spectrum of the AMO is also shown for reference.

for generating relative vorticity anomalies against processes that disperse and dissipate it. In the subtropical Atlantic (Fig. 5, middle), a zonally elongated band of anticyclonic RWS anomaly is collocated with the anticyclonic vorticity anomaly. Cyclonic RWS anomalies are found to the northwest and south of the anticyclonic RWS, and they are located near the cyclonic vorticity anomalies. The cyclonic forcing at around 30°N over the western North Atlantic, which is associated with converging anomalous wind, seems to play an important role in producing the northward propagating branch of the wave train, since it is located near the origin of the wave train as suggested by the diverging wave-activity flux. This cyclonic RWS is paired with wind divergence over the eastern portion of the subtropical North Atlantic sector. This divergent wind anomaly corresponds to a weakening of the convergence typically observed at the poleward edge of the Hadley cell. Strong RWS anomalies are also observed over the midlatitude and subpolar North Atlantic sector. These are mostly in quadrature with the relative vorticity anomalies and are mostly balanced by the advection of relative vorticity by the rotational component of the circulation [$\mathbf{u}_\theta \cdot \nabla(\zeta + f)$; not

shown]. These extratropical RWS are associated with vertical motions produced by the propagating wave train and thus may not necessarily indicate a true wave source.

The role of high-frequency transient eddies in the maintenance of the relative vorticity anomalies associated with NAO− is also assessed by regressing the feedback forcing by high-frequency eddies onto $\text{SVD1}_{\text{Z500ID}}$. The forcing is expressed as $-\nabla \cdot (\mathbf{u}_{\text{HF}} \zeta_{\text{HF}})$, where the subscript HF denotes the high-frequency component obtained through 10-day high-pass filtering. This feedback forcing (Fig. 5, bottom) is most effective in maintaining the vorticity anomalies in the extratropics, where climatologically the eddy-driven jet is located and the associated storm track is most active (not shown). The anticyclonic forcing is largest south of Greenland, where the anticyclonic anomaly associated with NAO− is located. The forcing also acts to maintain the cyclonic anomalies in the midlatitude North Atlantic and over Europe. This feedback forcing results from eddy–mean flow interactions internal to the atmosphere (Barnes and Hartmann 2010; Lorenz and Hartmann 2003) and anomalies in lower-tropospheric baroclinicity and circulation partly forced through ocean–atmosphere

FIG. 5. (top) Relative vorticity (color shadings) regressed on $\text{SVD1}_{\text{Z500ID}}$ (negative phase of interdecadal NAO), and the associated wave-activity flux (blue arrows; Takaya and Nakamura 2001) computed from the regressed geopotential height. A distance of 1° corresponds to a TN flux of $0.1 \text{ m}^2 \text{ s}^{-2}$. (middle) The corresponding interdecadal Rossby wave source (RWS) with color shading and the anomalous divergent wind component ($\mathbf{u}_d$) with green arrows. A distance of 1° corresponds to a speed of 0.02 m s^{-1} . (bottom) Feedback forcing by high-frequency eddies as vorticity tendency. All quantities are shown at the 300-hPa level. Red and grayscale shadings are used for, respectively, cyclonic and anticyclonic relative vorticity and feedback forcing. Statistically significant regions at the 95% level are contoured with magenta lines.

coupling (Peings and Magnusdottir 2014; Msadek et al. 2011; Peng et al. 2003).

The ocean–atmosphere interactions potentially involved in the interdecadal NAO variability are investigated by regressing 10-yr low-pass-filtered anomalies of SST, precipitation, and surface turbulent heat fluxes (sensible and latent heat fluxes combined; positive for upward fluxes) onto the $\text{SVD1}_{\text{Z500ID}}$ time series (Fig. 6). The analysis focuses on DJF because the primary goal is to explain the wintertime interdecadal evolution of the NAO and because air–sea coupling over the Atlantic sector is seasonally dependent (Gastineau and Frankignoul 2015;

Simpson et al. 2018). The interactions are assessed by comparing SST anomalies with turbulent heat fluxes at the ocean’s surface which is the prime way through which the ocean and atmosphere communicate and influence each other on interannual to longer time scales (Gulev et al. 2013). In addition to the analysis based on winter means that is shown here, we also repeated the regressions with annual means and found slightly weaker signals with similar geographic distributions. The regression analysis is performed for lags of -8 , -4 , and 0 (simultaneous) years to evaluate interactions that can lead to the formation of NAO-. At lag 0, the SST pattern is reminiscent of the

FIG. 6. Lag regression anomalies of (left) SST, (center) turbulent heat fluxes, and (right) precipitation on the SVD1_{Z500ID} (negative phase of interdecadal NAO) time series for lags of (top) -8 years and (middle) -4 years, and (bottom) the instantaneous regression. The regression of anomalous 10-m wind (black arrows) is overlaid on the turbulent heat fluxes. A distance of 1° corresponds to a speed of 0.07 m s^{-1} . Values that are significant at the 95% confidence level are contoured with magenta lines.

AMO (Deser et al. 2010; Yamamoto and Palter 2016; Kushnir 1994) with two pronounced lobes of warm anomalies over the subpolar and tropical North Atlantic. Although this pattern is similar to the AMO, we showed earlier that SVD1_{Z500ID} and SVD1_{SSVID} are only moderately correlated with the AMO index (Table 1). We also notice a resemblance with the SST tripole pattern associated with interannual NAO variability (Deser et al. 2010), although its midlatitude lobe is rather weak and confined to the western portion of the basin. The SST anomalies associated with SVD1 (NAO $-$) thus appear to be in between the typical signatures of the multidecadal AMO and the SST pattern associated with the interannual NAO.

Before the full development of NAO $-$ (lags -8 and -4 years), warm SST anomalies are observed over the western portion of the basin, including the Gulf Stream sector, where anomalous turbulent heat fluxes are upward. This is an indication of thermodynamic forcing by SST anomalies around the Gulf Stream onto the atmosphere. In fact, to the east of this anomalous heating, near-surface cyclonic anomalies associated with the southern center of action of NAO $-$ are developing, which is consistent with a steady linear response to shallow heating in the midlatitudes (Hoskins and Karoly 1981). By the peak of NAO $-$, the warm SST anomalies around the Gulf Stream diminish, resulting partly from

the persistent heat flux anomalies to the atmosphere, which at this stage, are maintained by the anomalous cool northerlies associated with the fully developed cyclonic anomalies. Contributions from anomalous oceanic heat transport, if any, cannot be evaluated here due to the lack of subsurface oceanic observational data for our analysis period. At lag 0, anomalous downward heat fluxes are collocated with the two warm Atlantic lobes, indicative of atmospheric driving of the SST anomalies. This pattern is reminiscent of the heat flux signature of the NAO (Deser et al. 2010; Delworth and Zeng 2016).

Leading to the development of NAO $-$ (lag -8 years), a zonal band of positive precipitation anomalies is observed over the tropical Atlantic, slightly north of the equator. This enhanced precipitation may result from a poleward displacement of the local ITCZ in the presence of the warm subtropical SST anomalies. Model experiments with prescribed tropical SST anomalies similar to the SVD1 signature produced precipitation anomalies comparable to the ones observed at lag -8 , which may yield an atmospheric response in the tropics and extratropics (Davini et al. 2015). This tropical signal, however, diminishes by the time NAO $-$ matures, when precipitation is enhanced in the subtropical central North Atlantic (around 30°N). We notice that the enhanced precipitation over this sector was not reproduced in the experiments by

FIG. 7. The (a) EAPE, (b) EKE, (c) CP, (d) CK, (e) CP_{HF} , (f) CK_{HF} , (g) CQ, and (h) CPK of subseasonal eddies (see section 2d) are regressed onto the SVD1_{Z500ID} (negative phase of interdecadal NAO) time series. Values that are significant at the 95% confidence level are contoured in magenta.

Davini et al. (2015). Our analysis suggests that this precipitation anomaly may be sustained in part by the enhanced turbulent heat fluxes from the Gulf Stream, to which latent heat flux contributes slightly more than half (not shown). These enhanced fluxes, combined with the cyclonic anomalies of NAO[−], yield an anomalous positive transport of moisture toward the eastern subtropical Atlantic where precipitation is enhanced. The precipitation anomaly implies an enhanced release of latent heat in the atmospheric column over the same sector, which likely forces ascent and upper-tropospheric divergence (cf. Fig. 5, middle). The anomalous divergent winds emanating from this region can yield anomalous vorticity advection in crossing the subtropical westerlies, to produce the cyclonic RWS (Fig. 5) and thereby force the stationary wave train.

d. Energetics of subseasonal eddies

This section investigates how NAO_{ID} variability modulates the energetics (see section 2d) of tropospheric subseasonal eddies. The energetics are first integrated

over the depth of the troposphere (1000–100 hPa) for each winter (DJF) and then locally regressed onto SVD1_{Z500ID}. Figure 7a shows that during the interdecadal NAO[−] the regressed EAPE of subseasonal eddies tends to increase over Canada, the eastern United States, and northern Europe, where SAT anomalies associated with SSV_{ID} are particularly large (Fig. 3a). In fact, lower-tropospheric temperature anomalies account mostly for the total EAPE. To explain the interdecadal modulations of EAPE by the NAO[−], we first examine how the baroclinic energy conversion (CP) from the background state to the subseasonal eddies is modulated (Fig. 7c). We observe a significant enhancement of CP over Alaska and Yukon, eastern North America, and western Europe. Among all the energy sources, changes in the efficiency of baroclinic energy conversion (CP^{eff}) contribute the most to the EAPE changes over the entire Northern Hemisphere (Fig. 8). The EKE of subseasonal eddies is also significantly modulated by the interdecadal NAO[−] with a large increase over the subpolar North Atlantic during NAO[−] (Fig. 7b). This EKE increase appears to

FIG. 8. Regression of hemispherically integrated energetics onto $SVD1_{Z500ID}$ (negative phase of interdecadal NAO). The regression of raw energy and energy conversion terms are shown with gray bars. Dark gray is used for statistically significant correlations at the 95% confidence level. The regression of the efficiency of energy sources and sinks is shown with blue borders. Significant changes in efficiency at the 95% confidence level are indicated with solid lines.

be supported by a weak increase in the barotropic energy conversion (CK) to subseasonal eddies over eastern North America (Fig. 7d). However, CK is reduced over more extensive areas over the North Atlantic sector, leading to only an insignificant reduction in the net conversion efficiency (CK^{eff}) during NAO– (Fig. 8). Instead, the net conversion efficiency from EAPE to EKE (CPK) increases significantly (Figs. 7h and 8), accounting for the notable EKE increase during NAO–. Meanwhile, the interdecadal modulations of the diabatic energy generation CQ are found negligible in the ERA-20C data (Figs. 7g and 8).

In comparison to the direct energy conversions from the background state to subseasonal eddies through CP and CK, the feedback forcing from high-frequency eddies plays a relatively minor role. The baroclinic feedback CP_{HF} from high-frequency eddies tends to be negative (Fig. 7e), as the associated heat flux is downgradient and thus acts as a damping of the thermal anomalies associated with subseasonal eddies. The net contribution thus leads to a significant net increase in damping efficiency (Fig. 8). During the interdecadal NAO–, the barotropic feedback (CK_{HF}) acts to enhance EKE significantly over the subpolar North Atlantic but dampens it over eastern Europe (Fig. 7f), leading to a small net enhancement of its efficiency (Fig. 8). Although the high-frequency eddy feedback is often considered important for the maintenance of subseasonal variability (Swanson 2002, and references therein), its role in the modulation of subseasonal variability by interdecadal NAO variability is relatively small.

The aforementioned analysis of energetics suggests that enhanced SSV during decades of NAO– may be

FIG. 9. CP composite differences between the positive and negative phases of $SVD1$ (featuring interdecadal NAO–) are calculated by keeping either (a) the background flow properties or (b) the eddy properties constant every year. For reference, the composite difference when both eddy properties and the basic state can change from year to year is shown in (c).

attributable mainly to the enhanced efficiency of CP. However, one important question that remains is whether the modulation of CP arises from interdecadal changes in eddy properties and/or in the basic-state circulation. To address this question, we evaluate the composite difference of CP and CK between the positive and negative phases of $SVD1_{Z500ID}$ (Fig. 9). For this exercise, the positive and negative phases are defined as the years when $SVD1_{Z500ID}$ rises above 0.5 and falls below -0.5 , respectively. To assess the role of eddies, a composite difference is evaluated by setting the basic-state terms (overbars) to their 1905–2005 averages, allowing only the eddy terms (primes) to change from year to year. Similarly, to evaluate the role of the background state, composites are performed by setting eddy covariance terms (primes) to their 1905–2005 average, allowing only the background terms (overbars) to change from year to year. These two assessments of CP are compared to another composite difference computed by allowing all the terms to change from year to year. The comparison reveals that the changes in CP are mostly due to the modulations in eddy properties (Fig. 9). Further analysis reveals that the subseasonal eddies tend to be more baroclinic during the decades of

FIG. 10. The column-mean correlation (a) between v' and T' and (b) between u' and T' is regressed onto the SVD1_{Z500ID} (negative phase of interdecadal NAO) time series. The regression is shown with black contours at intervals of 0.01 (solid and dashed lines for positive and negative values, respectively). Statistically significant regions at the 95% level are contoured with magenta lines. The climatology of the column-mean correlation is shown with color shadings.

NAO– (Fig. 10). Specifically, the negative correlation between u' and T' tends to be stronger around eastern Canada and Greenland, leading to a more efficient CP by acting on the eastward background temperature gradient (Fig. 10b). Likewise, the correlation between v' and T' tends to be more strongly positive over Europe and the eastern United States, resulting in a more efficient CP by acting on the southward background temperature gradient (Fig. 10a). Furthermore, more negative correlation between ω' and T' (not shown) is observed over eastern North America and to the west of Europe in agreement with the enhancement of CPK (Fig. 7h). We note, however, that these results do not necessarily minimize the importance of interdecadal changes in the background flow, which may modulate eddy structures and thereby indirectly affect their energetics.

The interdecadal modulation of subseasonal eddies by the NAO can also affect their propagation, or group velocity. It is here assessed with the wave-activity flux (Takaya and Nakamura 2001) evaluated from the bandpass-filtered 300-hPa anomalies, by assuming that subseasonal eddies are quasi-stationary Rossby waves.

The flux is calculated every day and then averaged over individual winters before it is regressed onto the SVD1_{Z500ID} time series. Figure 11 illustrates an enhancement of the eastward wave-activity flux of subseasonal eddies over the northeastern Pacific, North American, and Euro-Atlantic sectors, which is consistent with the enhanced eddy activity. A comparison with Fig. 7 reveals that the anomalous eastward flux tends to be observed in the vicinity and downstream of the sectors of enhanced subseasonal eddy energy sources (CP and CK).

e. Subseasonal wave trains and their interdecadal modulations

The typical structure of subseasonal eddies that induce SAT variability and its changes between the negative and positive phases of the interdecadal NAO are briefly investigated. Two reference locations are chosen where subseasonal SAT variability (Fig. 3a) and EAPE (Fig. 7a) are greatly modulated by interdecadal NAO variability: France (49°N, 2°E) and eastern Canada (53°N, 70°W). One-point regression maps are then

FIG. 11. The wave-activity flux (arrows; Takaya and Nakamura 2001) at 300 hPa for subseasonal eddies regressed onto the SVD1_{Z500ID} (negative phase of interdecadal NAO) time series. A distance of 1° corresponds to a flux of $0.7 \text{ m}^2 \text{ s}^{-2}$. Values that are statistically significant at the 95% confidence level are plotted with red arrows. Gray shading denotes $\text{CP} > 2 \times 10^4 \text{ J m}^{-2} \text{ day}^{-1}$ (Fig. 7c).

FIG. 12. Maps of one-point regression (shadings) and correlation (cyan contours) of subseasonal Z500 anomalies regressed onto reference time series of subseasonal SAT over (a),(c),(e) France (49°N , 2°E) and (b),(d),(f) eastern Canada (53°N , 70°W). Each of the patterns shown represents typical subseasonal anomalies obtained as an average over years when $\text{SVD1}_{\text{Z500ID}} > 1$ and $\text{SVD1}_{\text{Z500ID}} < -1$ (negative and positive phases, respectively, of interdecadal NAO). The contour intervals for Z500 and correlation are 5 m and 0.2, respectively, with solid and dashed lines representing positive and negative values, respectively. The Takaya–Nakamura (TN) flux is computed from the regressed Z500 field and shown with white arrows. A distance of 1° corresponds to a TN flux of $0.5 \text{ m}^2 \text{ s}^{-2}$. Red and blue dots indicate significant positive and negative differences, respectively, in the regressed local anomalies between years when $\text{SVD1}_{\text{Z500ID}} > 1$ and when $\text{SVD1}_{\text{Z500ID}} < -1$. Significance is assessed here with a two-sided t test at the 95% confidence level.

constructed by regressing subseasonal Z500 onto SAT time series at these reference locations. It is done separately for years when the normalized $\text{SVD1}_{\text{Z500ID}}$ index is larger than 1 ($\text{SVD1}+$) and smaller than -1 ($\text{SVD1}-$). To assess the origin and propagation of wave trains, the regression is repeated with time lags for which the Z500 anomalies lead the SAT anomalies (lags -3 and -6 days).

The left column of Fig. 12 shows one-point lag regression maps that represent typical subseasonal Z500 anomalies associated with warm SAT anomalies over France. At the peak time of the SAT anomalies (lag 0; Fig. 12e), a well-defined anticyclonic anomaly in Z500 is observed over France. As indicated by the wave-activity flux, the anomaly is part of a wave train propagating southeastward from a strong cyclonic anomaly over Greenland and Iceland into another cyclonic anomaly over the Middle East. The anomaly patterns at lags of -3 and 0 days are quite similar over Europe and the North Atlantic, indicating that the wave train is essentially quasi-stationary. At lag of -3 days (Fig. 12c), a

well-defined anticyclonic anomaly is observed over eastern Canada upstream of the cyclonic anomaly over Iceland. At lag of -6 days (Fig. 12a), the anticyclonic anomaly is stronger, while the anomaly over Europe is almost absent, indicative of group velocity propagation of the quasi-stationary Rossby wave train. The Z500 anomalies over Europe, Iceland, and northeastern Canada associated with the wave train tend to amplify significantly during decades of NAO $-$, which is consistent with the enhanced subseasonal eddy energy and fluxes highlighted in the preceding section. The differences are mostly due to amplitude changes rather than trajectory changes (not shown).

The same regression analysis is repeated for a reference SAT time series over eastern Canada (right column of Fig. 12). At lags of 0 and -3 days, the regressed subseasonal Z500 anomalies form a quasi-stationary wave train that appears to originate from the western subtropical Atlantic and propagate northward toward northeastern Canada before refracting southeastward toward Europe. At lag -6 , another wave train appears

FIG. 13. The influence of SVD1 on (a) frequency of cold extremes, (b) frequency of warm extremes, (c) wintertime mean temperature, and (d) subseasonal standard deviation of SAT, evaluated by regressing each of the four quantities onto SVD1_{Z500ID} (negative phase of interdecadal NAO). Significant correlations at the 95% confidence level are contoured in magenta.

to propagate over the eastern North Pacific toward Alaska. Again, the subseasonal Z500 anomalies over the Atlantic sector tend to be significantly stronger under the NAO[−] conditions, which is consistent with the increased EAPE and EKE observed over the sector (Fig. 7).

f. Decadal modulations of temperature extremes

Our final analysis is to investigate the impact of interdecadal NAO variability on surface weather extremes (Fig. 13). We define warm and cold extreme dates when the 10-day low-pass-filtered SAT (T_L) is above the 95th percentile and below the 5th, respectively, of the wintertime T_L distribution for all winters from 1900 to 2010. The use of T_L is to highlight subseasonal variability and prolonged extremes, such as heat waves and cold spells, that potentially have large socioeconomic impacts. As evident in Fig. 13a, the frequency of cold extremes if regressed on the normalized SVD1_{Z500ID} index (Fig. 13a) tends to increase over Europe and the eastern United States and decreases over northeastern Canada and south of Greenland. The corresponding changes in the frequency of warm extremes (Fig. 13b) are largely the opposite of those in cold extremes. Similar results are obtained for weather extremes defined with unfiltered SAT. These changes seem attributable primarily to changes in the winter-mean SAT (Fig. 13c),

since cold (warm) extremes are more (less) frequent where SAT is seasonally colder and vice versa.

Over Europe and northeastern North America, the enhanced subseasonal variability (Fig. 13d) during the interdecadal NAO[−] can act to broaden the range of subseasonal SAT fluctuations and thereby enhance the likelihood of extremes. Over regions like North Africa, where SVD1 exerts only a weak impact on winter-mean SAT, the enhancement of subseasonal SAT variability contributes to a slight increase in the likelihood of both warm and cold extremes.

4. Summary

This study documents, based on century-long reanalysis data, the dominant mode of covariability between the wintertime basic-state atmospheric circulation over the North Atlantic sector and subseasonal SAT variability on interdecadal time scales. Specifically, we have shown that SAT variance over northeastern North America and Europe is enhanced in decades when the NAO is in its negative phase (i.e., with a weakened and southward-deflected eddy-driven westerly jet over the North Atlantic). These decadal changes of the jet are due to a large-scale Rossby wave train that originates from the subtropical western North Atlantic and then propagates northward to the subpolar Atlantic

FIG. 14. Schematic diagram illustrating the key processes involved in the dominant mode of covariability described in this work (SVD1). (a) Interdecadal anomalies corresponding to the positive phase of SVD1 (or interdecadal NAO $-$). Because of their strong linearity, the opposite is true for the negative phase of SVD1 (or interdecadal NAO $+$), although the Rossby wave propagation remains the same. (b) Processes involved in subseasonal atmospheric variability whose SAT anomalies over Europe and Northeast America (color shaded) tend to strengthen under the positive phase of SVD1 (or interdecadal NAO $-$) shown in (a). The subseasonal processes in (b) subject to the interdecadal modulations include energy conversions from the winter-mean flow and subseasonal Rossby wave propagation, which tend to be enhanced (weakened) during interdecadal NAO $-$ (NAO $+$).

before being refracted southeastward toward Europe and the Middle East (Fig. 14a). Forced by enhanced precipitation in the central subtropical North Atlantic in its mature stage, this wave train, especially its northern portion, is projected strongly onto the negative NAO and maintained through feedback forcing from high-frequency transient eddies along the storm track (Feldstein 2003; Barnes and Hartmann 2010; Lorenz and Hartmann 2003; Lau and Nath 1991). Circulation anomalies associated with the NAO can also derive their energy directly from the basic-state flow (Martineau et al. 2020). The negative decadal NAO accompanies warm SST anomalies over the North Atlantic, whose spatial distribution is in between the AMO and the SST anomaly tripole typically associated with the monthly/interannual NAO (Deser et al. 2010). Anomalous winds associated with the decadal NAO are found to drive the SST anomalies

through anomalous sensible and latent heat fluxes once the NAO has fully developed.

In the developing stage of the negative decadal NAO, the associated cyclonic anomaly develops in the subtropics to the east of the warmer-than-normal Gulf Stream, where the cool anomalous northerlies act to increase upward sensible heat flux and surface evaporation. The latter acts to sustain the enhanced precipitation to the east, which in turn acts as a Rossby wave forcing to sustain the negative NAO, suggestive of a positive feedback. Unlike in some atmospheric GCM studies that have imposed SST anomalies similar to those associated with the interdecadal NAO (Peings and Magnusdottir 2016; Sutton and Hodson 2007; Davini et al. 2015), neither a significant shift of the ITCZ nor Rossby wave-activity flux coming from the tropics is observed in our analysis. This suggests that the

mechanisms forcing the atmospheric response may differ between the fully coupled atmosphere–ocean system and atmospheric GCMs. Another possible forcing of the negative NAO may be the modulation of midlatitude low-level baroclinicity by SST variability. It influences the growth of baroclinic eddies and their feedback on the eddy-driven jet (Peng et al. 2003; Msadek et al. 2011; Peings and Magnusdottir 2014), which is consistent with the eddy feedback observed in our analysis. The interdecadal variability of the NAO thus appears to be associated with a stationary wave train forced in the subtropics and maintained by high-frequency eddy feedback in the midlatitudes.

During decades of negative NAO, the increased subseasonal SAT variability over Europe and eastern North America is found to be a surface manifestation of enhanced EAPE aloft, which is supported by more efficient energy conversion from the baroclinic background flow over the North Atlantic sector (Fig. 14b). The decadal enhancement in the baroclinic energy conversion is found to result primarily from a stronger alignment between thermal and horizontal wind anomalies as decadal modulations in subseasonal eddy structure. This finding, however, does not minimize the role of the background interdecadal circulation anomalies, as they likely modulate the structural properties of the subseasonal eddies. A deeper understanding of how the structure of subseasonal eddies is modulated under the long-term changes in the background flow is thus required. During decades of the negative NAO, the EKE of subseasonal eddies is also enhanced over the subpolar Atlantic, which is found to arise primarily from a more efficient conversion from EAPE resulting from better alignment of thermal and vertical motion anomalies. In contrast, the net feedback forcing of subseasonal anomalies by high-frequency transient eddies is only of secondary importance in the interdecadal modulations of the energetics.

The mechanism proposed here to explain how interdecadal changes in the North Atlantic atmospheric circulation can modulate atmospheric variability on shorter time scales differs from the one proposed by Woollings et al. (2017). Their mechanism relies on changes in the trajectory and breaking of Rossby waves without consideration for changes in their amplitude. Here, we rather emphasize that the amplitude of subseasonal eddies is modulated on interdecadal time scales through changes in the efficiency of energy conversion from the background flow to subseasonal eddies. These two viewpoints are, however, not necessarily exclusive mutually, since it is likely that these two proposed mechanisms operate jointly.

Concerning changes in the frequency of SAT extremes, they rather strongly depend on the winter-mean

temperature response to the interdecadal NAO variability, which is similar to NAO-associated SAT fluctuations on shorter time scales (Hurrell et al. 2003). The negative phase of the interdecadal NAO is found to favor the occurrence of cold extremes over Europe and the eastern United States and warm extremes over northeastern Canada and Greenland.

Our maximum covariance analysis has extracted interdecadal changes in EAPE, EKE, energy conversion, and wave-activity fluxes of subseasonal anomalies not only over the North Atlantic sector but also over the eastern North Pacific and western North America, although in the latter regions the associated decadal modulations in the background state are rather modest (as illustrated schematically in Fig. 14a). More efforts should be dedicated in the future to understand the interbasin teleconnections (Kucharski et al. 2006; Honda et al. 2001, 2005b; Chafik et al. 2016) and their long-term modulations (Shi and Nakamura 2014) that may lead to a synchronized enhancement of subseasonal variability over the two sectors.

Modes of summertime covariability were also briefly investigated, though not discussed in detail. In contrast to its wintertime counterpart, the summertime covariance is spread more evenly among the modes (43% and 24% for the first two modes). The covariability is not apparently related to NAO-like variability, and the associated impact on SAT variability is more localized. This suggests that the modulation of subseasonal variability by interdecadal background flow changes is more complex in summer. Future work should investigate these summertime interactions in more detail.

The results presented here may have implications for future changes in subseasonal variability. Most CMIP5 models agree on a poleward shift of the North Atlantic jet stream throughout the year (Barnes and Polvani 2013) and such shift is already observed in summer and fall (Manney and Hegglin 2018). If this poleward trend in the jet stream axis projects onto the positive phase of the interdecadal NAO, a decrease of subseasonal variability may be expected as per the mode of covariability discussed in this work. Our findings further suggest that a reduction of the efficiency of baroclinic energy conversion may explain the projected reduction of waviness and the occurrence of blocking events as a response to climate change (Barnes and Polvani 2015).

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