Feedbacks of Sea Surface Temperature to Wintertime Storm Tracks in the North Atlantic

BOLAN GAN AND LIXIN WU

Physical Oceanography Laboratory, and Qingdao Collaborative Innovation Center of Marine Science and Technology, Ocean University of China, Qingdao, China

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ABSTRACT

In this study, the lagged maximum covariance analysis is performed on winter storm-track anomalies, represented by the meridional heat flux by synoptic-scale (2–8 days) transient eddies and sea surface temperature (SST) anomalies in the North Atlantic, which are both derived from reanalysis datasets spanning the twentieth century. The analysis shows significant seasonal and interannual coupling between storm-track and SST variations. On seasonal time scales, it is found that SST anomalies in the preceding early winter (November–December), which are expected to change the lower-tropospheric baroclinicity, can significantly influence storm tracks in early spring (March); that is, an intensification and slight northward shift of storm tracks in response to a midlatitude SST dipole, with a cold pole centered to the southeast of Newfoundland and a warm pole in the western subtropical Atlantic. This storm-track response pattern is similar to the storm-track forcing pattern in early spring, which resembles the dominant mode of storm tracks. At interannual time scales, it is found that the wintertime (January–March) storm-track and SST anomalies are mutually reinforced, manifesting as a zonal-dipole-like pattern in storm-track anomalies (with dominant negative anomalies in the downstream) coupled with a midlatitude SST monopole (with warm anomalies centered to the south and east of Newfoundland).

1. Introduction

Storm tracks, characterized by the intense activities of synoptic-scale transient eddies in the midlatitudes aloft (Blackmon et al. 1977), play a critical role in the climate system. These synoptic transients substantially affect not only local weather via the influence on precipitation, cloudiness, and winds, but also climate via the poleward transport of heat, moisture, and momentum and interaction with the large-scale mean flows (e.g., Trenberth and Hurrell 1994; Chang et al. 2002; Kug et al. 2010). Therefore, any systematic changes in the intensity and geographical positions of storm tracks would result in considerable changes in the extratropical daily weather and climate.

Observational evidence has indicated that the North Atlantic storm tracks exhibit pronounced seasonal-to-interdecadal variability. On seasonal time scales, storm tracks shift equatorward in the wintertime and are most intense in January as a result of the strongest tropospheric baroclinicity. This is in sharp contrast to the annual cycle of the North Pacific storm tracks, which undergoes striking midwinter suppression in intensity despite the strong baroclinicity (Nakamura 1992). On interannual time scales, it is found that storm tracks are closely related to the North Atlantic Oscillation (NAO), the dominant mode of interannual variability of atmospheric circulation over the North Atlantic, and the Arctic Oscillation (AO), the atmospheric internal variability over the extratropical Northern Hemisphere. During the positive phase of the NAO/AO, storm-track activity is intensified downstream (e.g., Nie et al. 2008; Rivière and Orlanski 2007; Wettstein and Wallace 2010). On decadal-to-interdecadal time scales, reanalysis data spanning the last half century indicate a significant transition from a weak to a strong storm-track regime around the mid-1970s, which is found to be linked with the NAO and AO variability (e.g., Chang and Fu 2002; Lee et al. 2012). The underlying mechanism, however, remains poorly understood.
understood. Given the large thermal inertia of the ocean, the oceanic variations are probably involved in the decadal-to-interdecadal variability of storm tracks.

Recently, storm tracks have been recognized to interact with the underlying ocean. On the one hand, the downward transport of mean westerly momentum mainly by the poleward transient heat flux acts to maintain the low-level westerly jet, which is collocated with the storm-track core and oceanic frontal zone (e.g., Lau and Holopainen 1984; Nakamura et al. 2004). The surface westerlies primarily drive oceanic gyres and influence sea surface temperature (SST) distribution through the processes of turbulent heat flux, Ekman advection, and entrainment [see a review by Kwon et al. (2010)]. On the other hand, the observed collocation of the surface westerlies, storm tracks, and SST frontal zones implies a potential influence of ocean on storm tracks. Indeed, recent modeling studies have demonstrated that the differential sensible heat supply from the ocean across a subarctic frontal zone is critical for restoring the near-surface baroclinicity, which is relaxed by the poleward transient heat transport and thus sustains the development of storm tracks (e.g., Nakamura et al. 2008; Taguchi et al. 2009; Sampe et al. 2010; Hotta and Nakamura 2011; Small et al. 2013). Hence the oceanic gyres driven by the surface westerlies associated with the storm-track activity contribute to the maintenance of oceanic frontal zones, which in turn fuels the growth of baroclinic transients, indicative of a positive feedback loop.

Observational and modeling studies also generally indicate that storm tracks are important for the air–sea coupling in the midlatitudes. On intraseasonal time scales, Ciasto and Thompson (2004) found a significant pattern of SST anomalies (SSTAs) in the Gulf Stream extension that precedes the winter NAO variability by ~2 weeks. Additionally, the basin-scale atmospheric circulation in early winter is found to display a barotropic NAO-like anomaly in response to a preceding summer horseshoe-like SST anomaly in the North Atlantic (e.g., Czaja and Frankignoul 2002; Ferreira and Frankignoul 2005). This SST anomaly pattern projects well on the tripole pattern generated by the NAO, thus indicating a positive feedback between the SST tripole and the NAO, which is confirmed by both observations and modeling studies (e.g., Peng et al. 2003; Pan 2005). Wu and Liu (2005) further suggest that this positive feedback acts as a key process to sustain decadal climate variability over the North Atlantic. In generating the atmospheric response in barotropic structure, storm tracks perturbed by SSTAs are revealed as a critical factor (Kushnir et al. 2002; Peng and Whitaker 1999; Peng et al. 2003; Taguchi et al. 2012). Specifically, the transient eddy forcing as a result of interactions between the diabatic-heating-forced anomalous flow and storm tracks transforms the initial baroclinic response into an equivalent barotropic one, suggesting the potential importance of the association between SSTAs and storm tracks in regulating the atmospheric responses.

Previous studies have revealed the importance of oceanic fronts associated with strong SST gradients on the climatological structure of storm tracks (e.g., Nakamura et al. 2008; Brayshaw et al. 2008; Woollings et al. 2010). In addition, the synoptic-scale transient eddies are found to covary with meridional shifts in the path of the Gulf Stream and Kuroshio–Oyashio Extension in winter (Joyce et al. 2009; Kwon and Joyce 2013). Nakamura and Yamane (2009) also identified the dominant anomaly patterns of the near-surface baroclinicity and the associated anomalies in SST and storm tracks. Overall, these findings imply a potential feedback of SST on storm tracks. The direct relationship between storm-track anomalies and SSTAs, however, has not been clearly identified, especially from observations. It motivates us to investigate the potential associations between the wintertime storm-track anomalies and SSTAs on seasonal and longer time scales. In this study, the investigation is based on the lagged maximum covariance analysis (MCA), which has been applied in the same way to examine such associations over the North Pacific (Gan and Wu 2013). On seasonal time scales, the North Pacific SSTAs in the preceding fall are found to significantly influence storm tracks in early winter, while at interannual time scales, the wintertime SSTAs and storm-track anomalies are mutually reinforced. Here, we find a significant influence of midlatitude SST dipole in the preceding early winter on storm tracks in early spring on seasonal time scales and a positive feedback between the midlatitude SST monopole and the zonal-dipole-like storm-track anomalies in the wintertime at interannual time scales.

The rest of the paper is arranged as follows. Section 2 presents the datasets and method used. The seasonal association between SST and storm-track anomalies in the cold season is described in section 3, followed by the wintertime association at interannual time scales in section 4. A summary and discussion is given in section 5.

2. Data and method

Atmospheric variables, including the meridional wind velocity, air temperature, surface wind stress, and net surface heat flux (the sum of shortwave, longwave, latent, and sensible heat fluxes), are obtained from the ensemble-mean fields of the Twentieth Century Reanalysis, version 2
Here, the temperature changes in the North Atlantic winter storm tracks (e.g., Czaja and Frankignoul 2002; Frankignoul and Sennéchael 2007). Here, the MCA is performed to investigate the relationship between storm tracks and SST in the North Atlantic. To enhance the signal-to-noise ratio, the domain for SST is chosen to highlight the area of strong meridional SST gradients, with the boundary set by 25.5°–60.5°N, 20.5°–80.5°W. The domain for storm tracks is 20°–70°N, 100°W–20°E. In fact, further inspection reveals that the present results are insensitive to the choice of analyzed region. To identify whether the MCA modes are meaningful, statistical significance is estimated using the Monte Carlo test, a nonparametric approach, in which MCA is repeated 100 times using the original SSTAs with storm-track time series randomly scrambled. The probability distribution function of the obtained 100 values of squared covariance (SC), squared covariance fraction (SCF), and correlation coefficients between the MCA time series is then constructed to rank the significance of the corresponding statistics.

In this study, the homogeneous map for SST and heterogeneous map for storm tracks, derived from the regression patterns against the SST time series of the MCA mode when SST leads storm tracks, are shown to investigate the influence of SSTAs on the storm-track activity. When studying the oceanic response to the storm-track variations, the heterogeneous map for SST and homogeneous map for storm tracks, which are projections on the storm-track time series of the MCA mode when storm tracks lead SST, are shown.

3. **Seasonal coupling between storm tracks and SST**

We first examined the seasonal association between SST and storm tracks during the cold season by applying the MCA as a function of time lag (months) to the monthly SSTAs and storm-track anomalies over the North Atlantic. It is found that the SC of the first MCA mode exhibits a strong seasonal dependence and an asymmetry of the lead–lag, as shown in Fig. 1. During the cold season, the SCs are large and broadly significant when storm tracks coincide with or lead SST, with a maximum in March at lag 0 and a secondary maximum in January at lag 4. This clearly indicates the predominant forcing of storm tracks on SST. In contrast, the SCs are much lower when SST leads storm tracks.
However, there are highly significant SCs when SST leads storm tracks in December and March by 2 and 3–4 months, respectively, indicating a potential influence of SSTAs on storm tracks. In fact, the significant SST-leading relation for storm tracks in December is not found in the better-sampled part of the 20CRv2 (1958–2008). Additionally, we have repeated the above analysis, but with the storm tracks measured by the standard deviation of 2–8-day bandpass-filtered geopotential height at 500 mb (1 mb = 1 hPa). It is found that the SCs are highly significant when SST leads storm tracks in March, whereas there are no significant SCs when SST leads storm tracks in December. Given the reduced robustness of the latter, here we mainly focused on the impact of SSTAs in the preceding early winter on storm tracks in early spring. Note that on seasonal time scales, no significant influence of SSTAs on storm tracks is found in the second MCA mode.

### a. Influence of SST on storm tracks

Figures 2a–d illustrate the maximum covariance patterns of SSTAs and storm-track anomalies for the corresponding first MCA modes with storm tracks fixed in March and SST in the preceding November, December, January, and February, respectively. Note that to examine whether there is a potential influence from the tropics, we extended the regression domain to include the tropical North Atlantic. When SST leads storm tracks, it is found that SSTAs in early winter can significantly influence storm tracks in the following early spring, manifesting as an intensification and slight northward shift of storm tracks in response to a midlatitude SST dipole with a cold pole centered to the southeast of Newfoundland and a warm pole in the western subtropical Atlantic. As seen in Figs. 2a–d, the covariance patterns, except for the western subtropical SSTAs, change little from lag −4 to lag −1, despite the fact that the corresponding SC at lag −2 is not highly significant. When SST leads by more than 1 month, however, the most significant impact of SSTAs on storm tracks in March occurs in the prior December, with an SCF of 57%. The corresponding SST time series at lag −3 exhibits a pronounced interannual variability, which is correlated with the storm-track time series, with the correlation of 0.34 significant at a 61% confidence level (not shown). The SST anomaly pattern shows cold anomalies dominant to the north of 40°N, with the maximum of −0.9°C off the southern coast of Newfoundland, accompanied by warm anomalies extending from the southeastern coast of the United States into the central basin (see colors in Fig. 2b). It resembles the first EOF of SSTAs (north of 20°N) in December, accounting for 22% of the total variance. In the tropical North Atlantic, there are cold anomalies with amplitude much weaker than the subpolar one, but comparable to the subtropical warm anomalies. In association with the prior SSTAs in December, storm tracks in March exhibit positive anomalies with the maximum located slightly to the north of its climatological peak, as shown in Fig. 3a, indicating an enhancement and slight northward shift of the storm-track activity (see contours in Fig. 2b). This storm-track response pattern resembles the first EOF of storm-track anomalies in March, accounting for 29% of the total variance (Fig. 3a). The maximum amplitude of atmospheric signal is 2 K m s⁻¹, implying a sensitivity of approximately 2.2 K m s⁻¹ °C⁻¹. At lags −4 and −2, the response patterns of storm tracks change little, although the corresponding time series are not highly correlated with the SST-leading time series, with the correlation significance of 60% and 44%, respectively.

To further determine which center of the SST anomaly pattern in December has the strongest influence on
storm tracks, we regressed the storm-track anomalies in March onto three SST time series in December, which are derived from the area-averaged SSTAs in the boxes centered on the main centers of action (see box in Fig. 2b). As seen in Figs. 3b–c, the storm-track anomaly pattern primarily associated with the cold center near 45°N and, to a lesser extent, the western subtropical warm center, resembles the corresponding MCA pattern, with the spatial correlation of 0.95 and 0.79, respectively. The maximum storm-track response is approximately 1.5 K m s^{-1} °C^{-1}. Compared with the other two centers of action, the eastern tropical cold center contributes little to the storm-track change, with a much lower spatial correlation of 0.46 (Fig. 3d). Thus, the midlatitude SST dipole in December is suggested to play a dominant role in affecting storm tracks after 3 months. Further discussion about the potential influence of the tropical Atlantic on storm tracks is given in section 5. Note that, given its independence from the MCA, this examination also suggests that the MCA results are robust.

According to Fig. 1, here the SST dipole in early winter is probably driven by storm-track variations. Indeed, the SST anomaly pattern derived from the simultaneous MCA mode for storm-track anomalies in November–December largely resembles the corresponding SST-leading pattern when storm tracks are fixed in March. This suggests that the midlatitude SSTAs are likely to be driven by storm tracks in the preceding early winter, then influencing storm tracks in early spring. Additionally, further inspection reveals that the storm-track anomaly pattern in December is associated with the NAO-like geopotential height anomalies at 850 mb. Cayan (1992) suggested that the dominant mode of monthly North Atlantic SSTAs (i.e., tripole mode) is closely related to changes in the local wind and surface heat flux, which is mostly controlled by the NAO.

Next, to understand the storm-track response to SSTAs, we diagnosed changes in the linear baroclinic instability of atmosphere, which provides a baroclinic source for the development of storm tracks. The baroclinicity is measured by the maximum Eady growth rate, defined as \( \sigma = 0.31gN^{-1}T^{-1} \partial T/\partial y \), where \( g \) is gravitational acceleration, \( T \) is the air temperature, and \( N \) is the Brunt–Väisälä frequency (Lindzen and Farrell 1980). In addition, given that a number of factors can influence the efficiency of eddies’ ability to tap into the baroclinicity of mean flow (e.g., Chang 2001), it is necessary to examine the local baroclinic energy conversion (BCEC). The BCEC from mean available potential energy (MAPE) to eddy available potential energy...
(EAPE) and from EAPE to eddy kinetic energy (EKE) are defined as follows:

\[
\text{BCEC}(\text{MAPE} \rightarrow \text{EAPE}) = -C_1 \left( \frac{P_0}{P} \right)^{R/c_p} \left( \frac{d\Theta}{dp} \right)^{-1} \left( \frac{\partial T}{\partial x} + \frac{\partial T}{\partial y} \right)
\]

\[
\text{BCEC}(\text{EAPE} \rightarrow \text{EKE}) = -C_1 (\omega T')
\]

where \( C_1 = (P_0/P)^{R/c_p} R/g \), and where \( R \), \( \omega \), \( \Theta \), and \( c_p (C_v) \) are the gas constant for dry air, vertical \( p \)-velocity, potential temperature, and specific heat capacity of dry air at the constant pressure (volume), respectively. The overbar signifies averaging over the individual winter months, and the prime denotes the synoptic-scale transients. More details are described in Cai et al. (2007).

These three quantities have been utilized in a range of studies on the storm-track variability and changes under global warming (e.g., Yin 2005; Lee et al. 2012).

Figure 4 shows the lagged regression maps of the maximum Eady growth rate and baroclinic energy conversion against the prior SST time series in December, with the corresponding climatology. The results suggest that the northward enhancement of atmospheric baroclinicity associated with the midlatitude SST dipole, with the strong cold anomalies north of the Gulf Stream extension in the preceding early winter, is presumably responsible for the northward intensification of storm tracks in early spring. As seen in Fig. 4 (top, left), the low-level Eady growth rate significantly increases along \( \sim 45^\circ \text{N} \) and decreases to the north and south. The positive regressions are located to the north of the climatological peak (Fig. 4 (top, right)) and collocated with the corresponding storm-track anomalies (Fig. 2b). A further inspection finds that changes in the baroclinic instability are primarily determined by the meridional temperature gradient changes. The strengthening of baroclinicity is clearly associated with an increase of eddy activities therein, as more available potential energy is likely to be tapped by the baroclinic eddies. This is indeed confirmed by the diagnosis of BCEC. The energy conversion from MAPE to EAPE (Fig. 4 (middle)) significantly increases over the Gulf Stream and its extension region; meanwhile, the energy conversion from EAPE to EKE (Fig. 4 (bottom)) increases, extending from the eastern United States to the United Kingdom. Both conversions show large regressions located slightly to the north of the maximum of the corresponding climatology.

Finally, we attempt to investigate the feedback of the storm-track response in early spring onto the ocean by regressing the net surface heat flux and surface wind stress in March onto the lag \( -3 \) SST time series in
December. It is found that both forcings on the ocean mostly act as positive feedbacks on SSTAs. The regression of the net surface heat flux shows that anomalous heat is transferred from the atmosphere into the ocean off the southern coast of Newfoundland, which damps the cold SSTAs therein (as seen by comparing Fig. 5b with the March SSTA pattern in Fig. 5a). However, there is an extensive region of anomalous cooling of ocean by atmosphere, thus reinforcing the cold SSTAs in the subpolar Atlantic. In addition, the regression of the surface wind stress shows a basin-scale anticyclonic anomaly over the subtropical North Atlantic, especially a strong eastward wind stress in the belt of 40–50°N (Fig. 5c). This is also clearly illustrated in the lagged regression map of the zonal wind at 850 mb (Fig. 6a). To further understand such features, we examined the corresponding regression map of the horizontal Eliassen–Palm vector $\mathbf{E}$, defined as $\mathbf{E} = \frac{1}{2}(\vec{u}^2 - \vec{u}^2)\mathbf{i} + (-\vec{u} \vec{v})\mathbf{j}$, where $\vec{u}'$ ($\vec{v}'$) and $\mathbf{i}$ ($\mathbf{j}$) denote the 2–8-day bandpass-filtered zonal (meridional) wind and the unit vector codirectional with the $x$ ($y$) axis, respectively. The $\mathbf{E}$ describes the forcing of transient eddies on the local mean flow, with the divergence and cyclonic curvature of $\mathbf{E}$ representing the eddy-induced acceleration of the westerly and southerly mean flow, respectively. As seen in Fig. 6b, $\mathbf{E}$ is significantly divergent over the northwestern subtropical Atlantic, coinciding with the enhanced westerly wind in Fig. 6a, which thus reflects its eddy-driven nature. As a result, the associated southward Ekman cold advection reinforces the cold SSTAs to the south of Newfoundland (Fig. 5d).

b. Storm-track forcing on SST

As seen in Fig. 1, the SC is largest at lag 0, with an SCF of 68%, reflecting the forcing of storm tracks on SST. The simultaneous covariance pattern shown in Fig. 2e suggests that the enhanced storm-track activity in March tends to induce a midlatitude SST dipole with strong cold and warm anomalies in the western subpolar and
subtropical Atlantic, respectively. Specifically, the storm-track pattern shows a northeastward tilt of a positive anomaly centered over the south of Newfoundland, resembling the dominant storm-track mode in March. This forcing pattern, however, bears a marked resemblance to the storm-track pattern responding to such an SST dipole in early winter (see Fig. 2b). The corresponding storm-track time series exhibit a pronounced interannual variability (not shown). The oceanic response shows a maximum cold SSTA of 0.7°C south of Newfoundland, accompanied by a maximum warm SSTA of 0.4°C off the southeastern coast of the United

![Image](https://example.com/image1)

**Fig. 5.** Lagged regression of the (a) SST, (b) net surface heat flux, (c) surface wind stress (N m$^{-2}$), and (d) Ekman advection in March onto the lag $-3$ SST time series in December. The contour interval is 0.1°C for (a), 3 W m$^{-2}$ for (b), and 5 W m$^{-2}$ for (d). Dashed and thick contours denote negative values and zero lines, respectively. Shaded areas and thick arrows indicate regression values significant at the 95% confidence level. Note that positive regression coefficients for the net surface heat flux signify the heat transferred from atmosphere into ocean, and vice versa.

![Image](https://example.com/image2)

**Fig. 6.** Lagged regression of the (a) zonal wind and (b) $E$ at 850 mb in March onto the lag $-3$ SST time series in December. (c),(d) As in (a),(b), but for the simultaneous regression against the lag 1 storm-track time series in March. Red (blue) contours signify the divergence (convergence) of $E$. The contour interval is 0.2 m s$^{-1}$ for (a), 0.15 m s$^{-2}$ (100 km)$^{-1}$ for (b), 0.5 m s$^{-1}$ for (c), and 0.3 m s$^{-2}$ (100 km)$^{-1}$ for (d). Dashed contours denote negative values. Shaded areas and thick arrows indicate regression values significant at the 95% confidence level.
States (Fig. 2e). It is similar to the dominant mode of North Atlantic SSTAs in March, accounting for 24% of the total variance. There are also cold SSTAs in the eastern tropical North Atlantic, with a magnitude much lower than the subpolar counterpart. When storm tracks lead SST, the covariance decays like the oceanic anomalies, which is clearly seen in the MCA mode of SSTAs and storm-track anomalies when the latter leads by 1 month (Fig. 2f).

To further investigate some of the processes responsible for the SSTAs induced by storm-track variations in early spring, we analyzed three heat budget terms that are directly associated with the storm-track forcing: the net surface heat flux $Q_{\text{net}}$, the Ekman advection $-\left(\rho C_p h V_{\text{Ek}} \cdot VT\right)$, and the vertical advection induced by Ekman pumping $-\left[\rho C_p h w_e (\partial T/\partial z)\right]$. In fact, the geostrophic advection $-\left(\rho C_p h V_{\text{geo}} \cdot VT\right)$ related to the storm-track variations is found to be generally insignificant, in spite of the large amplitude along the Gulf Stream, and, thus, this term is ignored here. Here $\rho C_p$ is the volumetric heat capacity of seawater, $T$ is the ocean temperature, $h$ is the mixed layer depth, $V_{\text{Ek}}$ ($V_{\text{geo}}$) is the horizontal Ekman (geostrophic) velocity, and $w_e$ is the Ekman-pumping velocity $[w_e = \text{curl}(\tau/\rho f)]$, where $\tau$ and $f$ are the horizontal wind stress vector and the Coriolis parameter, respectively. Note that positive $Q_{\text{net}}$ is defined to be downward (i.e., heat is transferred from atmosphere into ocean). The Ekman-pumping-induced advection in the wintertime is excluded in the deep convection regions, since the mixed layer depth therein exceeds 1000 m.

Figure 7 shows the simultaneous regression maps of the above three terms against the lag 1 storm-track time series in March. It is found that both the net surface heat flux and Ekman advection are responsible for the generation of the SST anomaly pattern. As seen in the comparison of Fig. 2e with Fig. 7, the surface heat flux forcings, especially the turbulent heat flux, mainly determine the cold SSTAs in both the subpolar and southeastern subtropical Atlantic and the warm SSTAs in the western subtropical Atlantic. However, the cold SSTAs in the slope water region slightly north of the Gulf Stream are primarily generated by the Ekman cold advection associated with the enhanced westerly wind along the belt of 40–60°N (Fig. 6c). Such intensification of low-level westerlies is driven by synoptic eddies, implied from the significant divergence of $E$ over the corresponding region (Fig. 6d). Further decomposition of the total Ekman advection $(V_{\text{Ek}} \cdot VT)$ into $V_{\text{Ek}}' \cdot VT'$, $V_{\text{Ek}} \cdot VT_0$, and $V_{\text{Ek}}' \cdot VT_0$, where the overbar denotes the climatological mean and the prime denotes the deviations, reveals that the advection of the mean SST by the anomalous Ekman current dominates over other components. This indicates a direct forcing of the cold SSTAs by the storm-track-induced Ekman advection.

Compared with the other two terms, the Ekman-pumping-induced vertical advection is smaller by one order of magnitude, marginally contributing to the warm SSTAs in the central subtropical Atlantic.

4. Interannual coupling between storm tracks and SST

To investigate the potential relationship between SSTAs and storm-track anomalies in the wintertime, we first examined the dominant modes of SSTAs, with removal of a third-order polynomial trend and the ENSO influence, and then regressed storm-track anomalies on
Here the wintertime anomalies are defined as the seasonal-mean anomalies in January–March (JFM), as the oceanic mixed layer is deepest in this period and thus SST is more persistent, which may enhance the oceanic influence on storm tracks. As seen in Fig. 8, the first and second EOFs of winter SSTAs in the North Atlantic (20°–70°N) exhibit two distinctive patterns. The first EOF, explaining 21% of the total variance, is characterized by a midlatitude dipole pattern with a center of action south and east of Newfoundland and a weaker center of opposite polarity off the southeastern coast of the United States (Fig. 8a). This SST mode resembles the SST anomaly pattern derived from the first MCA mode between SSTAs and storm-track anomalies in March (Fig. 2e). The second EOF, explaining 17% of the total variance, represents a monopole pattern with uniform polarity in the midlatitudes (Fig. 8b). The large SSTAs occur along the Gulf Stream (indicated by the tight isotherms of the JFM-mean SST for 3°, 6°, 9°, 12°, and 15°C are plotted in (b). The contours in (c),(d) signify the climatology of storm tracks in JFM.

Associated with these two SST modes, storm tracks exhibit distinct large-scale patterns. The regression of JFM-mean storm tracks onto the SST PC1 (Fig. 8c) shows positive anomalies nearly coinciding with its climatology position, indicating that an intensification of storm-track activity is associated with the warm SSTAs in the western subtropics and the cold SSTAs in the subpolar as well as the southeastern subtropical Atlantic. This is consistent with the instantaneous MCA results of storm-track anomalies and SSTAs stratified by calendar month in winter (see Fig. 2e as an example). Associated with the warm SST monopole, storm tracks exhibit northeasterward-tilted positive anomalies in the northern flank of the corresponding climatological center and negative anomalies along the southern flank and downstream region (Fig. 8d). This suggests a more pronounced northeasterward tilt of storm tracks.

To further investigate whether the aforementioned association between the wintertime storm-track anomalies and SSTAs is coupled at interannual time scales, we applied the MCA as a function of time lag (years) to the JFM-mean SSTAs and storm-track anomalies, with the ENSO influence removed, in the North Atlantic. As seen in Fig. 9, the SC, SCF, and correlation coefficients of the first MCA mode show a coherent significance at lag 0, whereas no significant statistics are detected when SST leads storm tracks. Indeed the instantaneous covariance pattern of SSTAs and storm-track anomalies, accounting for 56% of the SC, is similar to the pattern shown in Figs. 8a and 8c. The second MCA mode, however, shows consistent significance of the three
statistics identified from lag −1 to 1, indicating that at interannual time scales, changes in winter SST are coupled with the storm-track variations.

Figure 10 illustrates the maximum covariance patterns of SSTAs and storm-track anomalies, for the corresponding second MCA modes at lag −1, 0, and 1. When SST leads storm tracks, an east–west dipole-like structure of storm-track anomalies is found to be associated with a monopole-like pattern of midlatitude SSTAs in the previous winter. As seen in Fig. 10a, the SST anomaly pattern has warm anomalies in nearly the entire North Atlantic, with two peaks of ~0.7°C occurring to the south and east of Newfoundland. The corresponding SST time series (not shown) exhibits pronounced decadal-to-multidecadal variability with a spectral peak at ~40 years, which is closely related to the Atlantic multidecadal oscillation index (i.e., the 10-yr running mean of the detrended Atlantic SSTAs poleward of equator; Enfield et al. 2001), with a high correlation of 0.62 (after 10-yr low-pass filtering the SST time series). Associated with this SST mode 1 year later, storm tracks show negative anomalies to the east of the warm SSTA’s center and positive anomalies over the western subpolar Atlantic, with the maximum of ±0.6 K m s⁻¹, implying a sensitivity of approximately 0.9 K m s⁻¹ °C⁻¹. This storm-track anomaly pattern is found to account for ~13% of the storm-track anomaly variance in JFM. In fact, by regressing the JFM-mean storm-track anomalies 1 year later onto the SST time series derived from the area-averaged SSTAs in the box of 40.5°–50.5°N, 37.5°–59.5°W, such negative anomalies are found to be statistically significant and dominant over the positive ones. Furthermore, the lagged regression of the maximum Eady growth rate at 850 mb onto the lag −1 SST time series shows that the lower-tropospheric baroclinicity significantly weakens in the belt of 40°–50°N, which is presumably responsible for the downstream weakening of storm-track activity (Fig. 11a). The corresponding lagged regression map of the local baroclinic energy conversion also displays significant negative anomalies in the central-eastern North Atlantic, indicating that less available potential energy is tapped by the baroclinic eddies and thus results in the weakening of storm tracks (Figs. 11b,c).

For the storm-track forcing on SST, the instantaneous MCA result (Fig. 10b) shows positive storm-track anomalies in a northeastward direction, dominant over the downstream negative anomalies and warm SSTAs across the subtropical Atlantic, with large anomalies occurring along the oceanic frontal zone. This mode largely resembles the second EOF of the JFM-mean SSTAs and the associated storm-track anomalies (Figs. 8b,d).
Additional examination reveals that this covariance pattern mainly reflects the forcing of the positive pole of the dipolar storm-track anomalies, as shown in Fig. 10a, onto the North Atlantic Ocean. The influence of the negative pole on SST can be detected in the storm-track-leading situation as follows.

As seen in Fig. 10c, the storm-track anomaly pattern shows maximum positive (negative) anomalies with a magnitude of 2 K m s\(^{-1}\) along the northwestern (southeastern) flank of the climatological storm tracks, which resembles the third EOF of storm-track anomalies in JFM, accounting for 11% of the total variance. This mode depicts a deflection of storm tracks toward Greenland and weakened eddy activities over the eastern North Atlantic. The corresponding storm-track time series at lag 1 exhibits pronounced interannual-to-decadal variability, with significant spectral peaks at around 13–17 years and a suggestion of a peak at the multidecadal period as well (not shown). This is similar to the power spectrum of the storm-track time series for the SST-leading situation. The associated SST anomaly pattern shows basin-scale warm SSTAs in the midlatitudes, with a maximum of 0.3°C centered near 50°N, 40°W. A further
inspection of the simultaneous regression of three heat budget terms onto the lag 1 storm-track time series suggests that the downward heat flux in the central part of the northern North Atlantic is mainly responsible for the warm SSTAs therein (Fig. 12a). In addition, the warm SSTAs south of Newfoundland are mainly attributed to the anomalous Ekman warm advection along ~42°N associated with a basin-scale anticyclonic wind stress anomaly centered at 55°N, 25°W (Fig. 12b). The Ekman-pumping-induced vertical advection is found to have little contribution (Fig. 12c). Here, it is worth noting that the storm-track forcing pattern is similar to the pattern responding to the monopole SST mode, as seen in the comparison of Fig. 10a with Fig. 10c. This suggests that at interannual time scales, the zonal-dipole-like storm-track anomalies, with dominant negative anomalies in the downstream, are mutually reinforced with the basin-scale warm SSTAs in the midlatitudes during boreal winter.

5. Summary and discussion

The relationships between the wintertime SSTAs and storm-track anomalies over the North Atlantic at seasonal and interannual time scales are investigated based on the lagged MCA. The MCA results suggest that the basin-scale storm-track anomalies are significantly associated with distinctive patterns of SSTAs at different time scales.

On seasonal time scales, it is found that SSTAs in the preceding early winter can significantly influence storm tracks in early spring, such as an intensification and slight northward shift of storm tracks in response to a midlatitude SST dipole with a cold pole centered to the southeast of Newfoundland and a warm pole in the western subtropical Atlantic. The maximum response magnitude is approximately 2.2 K m s\(^{-1}\) °C\(^{-1}\). Further analyses of the maximum Eady growth rate and baroclinic energy conversion suggest that the northward strengthening of lower-tropospheric baroclinicity associated with such an SST dipole in the preceding early winter is presumably responsible for the northward enhancement of storm-track intensity in early spring. Through composite analysis, Nakamura and Yamane (2009) also found that the enhanced near-surface baroclinicity in early spring is associated with a dipole-like SST anomaly in the three prior months, similar to the aforementioned one.

As for the feedbacks of North Atlantic SSTAs to the large-scale atmospheric circulation, we explored the early spring atmospheric response to the midlatitude SST dipole in the preceding early winter. As seen in Fig. 13, the atmospheric response significantly exhibits an equivalent barotropic structure with the large-scale cyclonic anomalies centered to the west of Greenland and the anticyclonic anomalies over the subtropical North Atlantic, resembling the NAO pattern. This dipolar pattern also bears some similarity to the response pattern of geopotential height to the North Atlantic Ocean variability (Gastineau et al. 2013) and the anomalous synoptic-eddy forcing associated with the North Atlantic SST tripole (Peng et al. 2003). In fact, Czaja and Frankignoul (2002) showed a significant SC of the first MCA mode between the North Atlantic (20°–70°N) SSTAs in October–December (OND) and the 500-mb geopotential height anomalies in February–April (FMA), Fig. 12. Simultaneous regression of the (a) net surface heat flux, (b) Ekman advection, and (c) Ekman-pumping-induced vertical advection onto the storm-track time series derived from the second MCA mode at lag 1. The contour interval is 5 W m\(^{-2}\) for (a), 3 W m\(^{-2}\) for (b), and 1 W m\(^{-2}\) for (c). Dashed and thick contours denote negative values and zero lines, respectively. Shaded areas indicate regression values significant at the 95% confidence level. Note that positive regression coefficients for the net surface heat flux signify the heat transferred from atmosphere into ocean, and vice versa.
in which storm tracks probably play an important role (Peng and Whitaker 1999).

Given the tropical Atlantic impact on the extratropical atmosphere (e.g., Drévillelon et al. 2003; Peng et al. 2005) and significant cold SSTAs shown in the eastern tropical North Atlantic (Fig. 2b), we further examined the role of the tropical Atlantic in influencing storm tracks by conducting the lagged MCA between the monthly Pan-North-Atlantic (0°–60°N)/tropical North Atlantic (0°–20°N) SSTAs and midlatitude (20°–70°N) storm-track anomalies, with the ENSO influence removed. As seen in Fig. 14 (top), the SCs of the first MCA mode based on the Pan-North-Atlantic SSTAs are highly significant when SST leads storm tracks fixed in February in addition to December and March. Such a significant SST influence on storm tracks in February probably results from the tropical North Atlantic, which is clearly implied from the MCA result derived from the

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**Fig. 13.** Lagged regression of geopotential height anomalies at (left) 850 and (right) 250 mb in March onto the lag −3 SST time series in December. The contour interval is 2 m. Dashed contours denote negative values, and shaded areas indicate regression values significant at the 90% confidence level.

**Fig. 14.** As in Fig. 1, but for the first MCA mode between the Pan-North-Atlantic (0°–60°N)/tropical North Atlantic (0°–20°N) SSTAs and midlatitude (20°–70°N) storm-track anomalies.

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**SC (× 10^3) of MCA1 <V'T', SST(0°–60°N)>**

**SC (× 10^3) of MCA1 <V'T', SST(0°–20°N)>**

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tropical North Atlantic SSTAs and midlatitude storm-track anomalies [Fig. 14 (bottom)]. Further inspection reveals that the cold SSTAs in the eastern tropical North Atlantic in November and December can significantly weaken the upstream storm-track intensity in February. This is in sharp contrast to the impact of the midlatitude SST dipole in early winter on storm tracks in March. It is worth noting that when the tropical North Atlantic SSTAs lead storm tracks in March by 1–4 months, no significant SCs are detected. This suggests that the significant SCs identified at lag $-3/-4$ with storm tracks fixed in March in Fig. 1 primarily reflect the response of storm tracks to the midlatitude SSTAs rather than the remote forcing of the tropical Atlantic. Additionally, given the significant SCs detected in Fig. 14 (bottom) with SST-leading storm tracks in December, we speculate that the similar SCs shown in Fig. 1 may arise from the tropical Atlantic influence.

Here the detected 3-month lag mainly reflects the persistence of SSTAs rather than the occurrence of the SST forcing in advance, as suggested by Czaja and Frankignoul (2002). We further examined the autocorrelations of the monthly storm-track and SST time series from November to May, which is shown in Figs. 15a–b plotted as a function of the calendar months. The wintertime SSTAs clearly exhibit large persistence of 3 months, at least, which explains why the influence of SSTAs on storm tracks that is likely to happen in early spring can be detected by using SSTAs in December (Fig. 15b). Indeed, the cold pole of SST dipole in early winter persists into the early spring (see Fig. 5a), which is expected to change the meridional air temperature gradient and thus influence storm tracks. For storm tracks, however, the correlation between two neighboring months is only about 0.1, indicating that winter storm-track anomalies cannot persist into the next month (Fig. 15a). This suggests that the significant SC at lag $-1$ with storm tracks fixed in March probably reflects the influence of SSTAs on storm tracks rather than a remnant of the storm-track forcing on SST. An issue
also arises regarding the lower and less significant SC for SST leading the March storm tracks by 2 months (Fig. 1). We speculate that this may result from the reduced signal-to-noise ratio for the midlatitude SST dipole in January. Further examination finds that the mixed layer depth averaged over the western North Atlantic in January is 50% higher than that in December. Such deepening of the mixed layer may involve anomalies from subsurface, which may mask the signals affecting storm tracks. Indeed, the SST dipole in December accounts for 31% of the SST anomaly variance, versus 27% for the SST pattern in January.

Note that further analysis of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data and the better-sampled part of the 20CRv2, with the temporal coverage of 1958–2008, mostly supports the results of SSTAs’ feedbacks to storm tracks in early spring, albeit with a much broader response of storm tracks. During the second half of the twentieth century, however, the SCs of the first MCA mode display high significance when SST leads storm tracks in March by 1–2 months. This suggests that the significant MCA mode at lag $-3/4$ identified here may mainly reflect the dynamics in the first half of the twentieth century, and the response of storm tracks to SSTAs may be potentially modulated by changes of the background state.

For the forcing of storm tracks on SST, it is found that the enhanced storm-track activity in early spring tends to induce strong cold and warm SSTAs in the western subpolar and subtropical Atlantic, respectively. Furthermore, the generation of SSTAs is attributed to both the net surface heat flux and Ekman advection associated with the storm-track variations. Note that the storm-track forcing pattern in early spring is similar to the pattern responding to the SST dipole in the western North Atlantic in the previous early winter, which may help to sustain SSTAs. This is in sharp contrast to the seasonal coupling between SSTAs and storm-track anomalies in the North Pacific, such that the basin-scale cold SSTAs in the subpolar region in the preceding fall can northward enhance the storm-track activity in early winter, whereas a meridional shift of storm tracks tends to force a horseshoe-like SST anomaly in early winter (Gan and Wu 2013).

At interannual time scales, a positive feedback is found between the zonal-dipole-like storm-track anomalies (with dominant negative anomalies in the downstream) and the midlatitude SST monopole (with warm anomalies centered to the south and east of Newfoundland) in the wintertime (JFM). The maximum response of storm tracks to the SST monopole is approximately $0.9 \text{K m s}^{-1} \text{C}^{-1}$. This positive feedback may enhance the decadal climate variability in the North Atlantic and may also contribute to the interannual-to-decadal variability of the zonal-dipolar storm tracks in winter. Additional examination reveals that the negative anomalies of storm tracks in response to such warm SSTAs are closely related to the weakening of lower-tropospheric baroclinicity, which is also dynamically consistent with the fact that the baroclinic eddies tap less available potential energy from the mean flow.

Here the SST monopole in winter affecting storm tracks in the following winter is likely to be attributed to the large winter-to-winter persistence of SSTAs, especially in the two centers located to the south and east of Newfoundland. Indeed, further examination of the autocorrelations of the JFM-mean SST and storm-track time series finds that the winter SSTAs show large interannual persistence, whereas the winter storm-track anomalies display little persistence (Figs. 15c,d). It seems that the persistent SSTAs cannot be interpreted by the “reemergence mechanism” which involves the seasonal variation of the mixed layer depth (e.g., Alexander and Deser 1995), since the persistent SST centers appear not to align with the two reemergence areas in the North Atlantic identified by Hanawa and Sugimoto (2004). Thus, the oceanic processes, such as the heat transport by the Gulf Stream, are speculated to contribute to the large persistence of the SST monopole. In fact, Kwon and Joyce (2013) recently found that a northward shift of the Gulf Stream, associated with warm SSTAs extending from the Cape Hatteras to the east of the Grand Banks, can induce a weakening of the synoptic transient eddies in the downstream region 1 year later. This is consistent with the present result of SSTAs’ feedback on storm tracks at interannual time scales. Further modeling experiments, however, are needed to understand the mechanisms underlying the positive feedback identified here. It is also worth noting that the impact of winter SSTAs on storm tracks in the North Atlantic is merely detected when SST leads by 1 year, which is shorter than that detected in the North Pacific: that is, 3 years (Gan and Wu 2013). This may suggest a larger impact of the North Pacific Ocean on storm tracks and/or a larger persistence of SSTAs in the Kuroshio–Oyashio Extension region.

At seasonal and interannual time scales, the wintertime storm tracks exhibit a different anomaly pattern in response to the North Atlantic SSTAs. Further inspection reveals that the different response patterns did not arise from the different time lags or JFM-mean versus monthly anomalies used. We speculate that such difference may be due to the SST anomaly mode dominated by different variability. For the seasonal coupling between storm tracks and SST, the SST anomaly pattern has a pronounced interannual variability, which is likely
to be a passive response to the atmospheric forcing. However, for the interannual coupling, the SST anomaly pattern shows significant decadal-to-multidecadal variability, which may be induced by changes in the ocean dynamics and thus has large year-to-year persistence.

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