Impacts of mesoscale sea surface temperature anomalies on the meridional shift of North Pacific storm track

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Abstract
In this study, an index for the variability of mesoscale sea surface temperature (SST) anomalies over the Kuroshio and Oyashio confluence region (KOCR) is proposed based on the high-resolution SST data set. The positive phase of the new index indicates that the mesoscale SST anomalies are strong and feature double peaks of variance within the KOCR, and the negative phase of the index denotes weak mesoscale SST anomalies and a broader single peak of variance. Composite analysis is conducted on the relation between the variability of mesoscale SST and the North Pacific storm track. The mechanism behind the relation is then investigated. We find that the storm track is shifted southwards (northwards) during the positive (negative) phase of the index. The variability of mesoscale SST impacts first the turbulent heat fluxes out of the ocean, changing the near-surface baroclinicity with the large-scale zonal wind. The baroclinic energy conversion resembles the baroclinicity anomaly, modulating the anomaly of the storm track along the southern side of its climatological peak.

KEYWORDS
air–sea interaction, baroclinic energy conversion, mesoscale sea surface temperature anomaly, North Pacific storm track

1 | INTRODUCTION

In the western boundary current and its extension, such as the Kuroshio Extension (KE), energetic ocean mesoscale eddies and fronts contribute greatly to mesoscale sea surface temperature (SST) variability (Chelton et al., 2004; Xie, 2004; Small et al., 2008; Bryan et al., 2010; Neill et al., 2012; Bishop et al., 2017). The mesoscale SST anomaly (SSTA) is tightly coupled with the atmosphere (Minobe et al., 2008; Ma et al., 2016b). Previous studies using high-resolution satellite observations and numerical models showed that it is the ocean that drives the atmosphere at the mesoscale (e.g., Bryan et al., 2010; Chelton and Xie, 2010; Frenger et al., 2013; Ma et al., 2016a; Bishop et al., 2017).

Ocean fronts can significantly modulate the strength and location of storm track (Nakamura et al., 2004; Joyce et al., 2009; Taguchi et al., 2012; Small et al., 2014; O’Reilly and Czaja, 2015). Using the classic baroclinic theory, Nakamura et al. (2004; 2008) revealed that the storm track strengthens along the baroclinic zone. Taguchi et al. (2009) suggested that the surface storm track could shift meridionally, along with the decadal displacement of the subarctic frontal zone (SAFZ). Small et al. (2014) showed that the storm track response to the smoothed SST is diminished by 10–20%. O’Reilly and Czaja...
(2015) indicated that storm track peak displays a zonal shift associated with the change of KE front strength. Besides mesoscale oceanic fronts, the influence by oceanic eddies has also been found to be important, which is not only confined within the boundary layer, but also penetrates the free troposphere, leading to the variability of the storm track (Wang and Liu, 2015; Ma et al., 2015a; 2015b; 2017). Chelton and Xie (2010) revealed a positive correlation between mesoscale SSTAs (MSSTAs) and surface winds, which showed the surface wind speed strengthens over warm eddies and weakens over cold eddies. After removing the MSSTA by spatial smoothing, the storm track shifts towards the maximum SST gradient (Small et al., 2014; Piazza et al., 2016). Ma et al. (2015b) showed that the MSSTA variability in the KE could remotely impact the winter rainfall over the eastern northern Pacific and enable the diabatic energy conversion between latent heat energy and transient eddy energy, enhancing the storm track. More recently, Ma et al. (2017) extended their study and further explored the important role of MSSTAs on simulating the North Pacific storm track. They suggested that there is not only a local reduction in cyclogenesis but also a southwards shift of downstream storm track after removing the MSSTAs in the KE region. However, the mechanism of MSSTA-driven air–sea interactions is still controversial.

Two mechanisms, through which the MSSTAs affect the atmosphere, have been proposed. One is the pressure adjustment mechanism (PAM) by Lindzen and Nigam (1987), in which the low sea level pressure (SLP) anomaly is induced by warm SSTA, forcing a convergence towards the anomaly centre through pressure gradient. The other is the vertical mixing mechanism (VMM) by Wallace et al. (1989), in which the eddy-induced SSTAs can significantly adjust surface wind through vertical momentum mixing; this process is strongly related to the atmospheric stability. Putrasahan et al. (2013) has examined the scale dependence of the two mechanisms, VMM and PAM. Their results showed that both VMM and PAM are active on the mesoscale. Furthermore, the relative importance of the two mechanisms was found to change with timescale. Liu et al. (2013) and Chen et al. (2017) suggested that the VMM is clearer at the synoptic timescale, while the PAM needs longer adjustment time of about 1 month. However, the MSSTAs over the Kuroshio and Oyashio confluence region (KOCR) and the meridional shift of storm track have not been investigated comprehensively so far. Which mechanism can explain the meridional shift of storm track is still an open question.

The purpose of the present paper is to investigate the variability of MSSTA over the KOCR and its possible influence on the North Pacific storm track. Two scientific questions are addressed: (a) What is the relationship between the variability of MSSTA and storm track? (b) What is the underlying dynamics contributing to the relationship? The “storm track” in this paper refers to the North Pacific storm track, unless stated otherwise. This paper is organized as follows. In section 2, we describe the data and define an index of MSSTAs. In section 3, we show the relationship between MSSTA and storm track. In section 4, we reveal the largescale atmospheric circulation response to MSSTAs. In section 5, we explore the mechanisms of storm track response to the variability of MSSTAs. In section 6, conclusions of the paper are given.

2 | DATA AND METHOD

2.1 | Data

In this study, we use the daily SST with the horizontal resolution of 0.25 × 0.25° from the National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation Sea Surface Temperature (OISST). The SST data combine satellite observations and in situ observations since 1981 (Reynolds et al., 2007). The SST variation due to mesoscale eddies can be partially resolved at this resolution. The mesoscale signal of SSTA can be derived from this data set, and the method used will be discussed below.

The atmospheric variables, including mean SLP, geopotential height, air temperature, zonal wind, meridional wind and vertical velocity, are all obtained from the NCEP/NCAR reanalysis on a 2.5 × 2.5° grid (Kalnay et al., 1996). Daily surface heat fluxes are provided by the objectively analysed air–sea fluxes (OAFlux; Yu et al., 2008) with the spatial resolution of 1 × 1°.

The period from 1985 to 2005 is chosen. The cold season here is defined from October to February of the following year, to take late-autumn into consideration.

2.2 | Index of MSSTA

To separate the MSSTA from large-scale SST signals, a 5 × 5° spatial boxcar filter is employed on the daily SST, following the studies of Itoh and Yasuda (2010) and Putrasahan et al. (2013), in which they showed most mesoscale eddies have diameters smaller than 5°. Figure 1a–c shows the climatology of the unfiltered, low-pass-filtered (large-scale) and high-pass-filtered (mesoscale) SSTA in the cold season, respectively. Most of the MSSTA features are confined within a region (142°–175°E, 33°–45°N), denoted by the red box. Also, the region is characterized by large variance of MSSTA, which contributes about half (49%) of the total variance. In fact, the substantial fraction of the monthly mean SSTA variability is intrinsic to the ocean in the KOCR, other than triggered by atmospheric forcing (Smirnov et al., 2015). Furthermore, we have tested the method to separate the meso- and large-scale signals in SST
and compared with other method. Ma et al. (2015a; 2015b; 2017) used a Loess spatial filter with $15 \times 5^\circ$ cut-off wavelength to remove the mesoscale SST whose half-power wavelength is approximately 900 km, while our results is about 500 km. It is also evident that the SSTAs with wavelength longer than 1,000 km are removed efficiently through high-pass filtering (Figure 1d). Therefore, we believe the $5 \times 5^\circ$ spatial boxcar filter did a good job in separating signal of two scales.

To quantitatively measure the variability of MSSTAs, we define a MSSTA index as follows. First, the daily mean sea surface temperature ($T$) is separated into the spatial averaged ($< T >$) and the mesoscale components ($T^*$) by using the $5 \times 5^\circ$ high-pass boxcar filter. Therefore, the daily mesoscale anomalies of SST could be expressed as: $T^* = T - < T >$. Second, the variance of daily MSSTA ($T^* T^*$) is calculated at every grid point. Third, the area and monthly average of the daily variance ($\overline{\sigma_{T^* T^*}}$) within KOCR is obtained, where $\overline{[\ ]}$ and ($\cdot$) stand for the spatial mean within KOCR and each month, respectively. The values for each month can be expressed as: $V_i = \overline{\sigma_{T^* T^*}}$, “$i$” from 1 to $N$, where $N = 110$ months. Finally, the $\bar{V}$ and the standard deviation ($S = \sqrt{\frac{1}{N} \sum_i (V_i - \bar{V})^2}$) can be computed, and the MSSTA index is defined as $M_i = \frac{V_i - \bar{V}}{S}$.

Figure 2 shows the normalized monthly mean time series of MSSTA index by the period mean and its standard deviation. Positive (negative) MSSTA index denotes strong (weak) variability of MSSTAs, which include both mesoscale eddies and the meanders of fronts. The time series of MSSTA index clearly shows the variability on the inter-annual and decadal timescales.
The positive phase is defined when the MSSTA index is greater than 1.0, and the negative phase refers to months when MSSTA index is less than −1.0. According to this criterion, 20 months are selected for the positive phase and 17 for the negative phase. The composites are calculated by averaging the SSTA during 20 positive-phase months and 17 negative-phase months (Figure 3a,b), respectively. Figure 3 also shows the composites of the monthly variance of MSSTAs for positive phase and negative phase (Figure 3c,d). It is evident that the variance is stronger during positive phase than that during negative phase, which also indicates stronger SSTA during positive phase. Besides, we can find out that there is a marked difference in the probability density function distribution with increased (decreased) occurrence of stronger anomalies during the positive (negative) phase (Figure 3e), which indicates that the variability of MSSTAs is more intense during the positive phase.

Figure 3f shows the zonal mean MSSTA variance averaged between 142°E and 175°E. During the positive phase, there are dual local maxima of variance, one near 36°N and the other around 41°N with maximum values exceeding 1°C. The ocean fronts may contribute to this pattern. Masunaga et al. (2015) showed that the equatorwards SST gradient also has dual local maxima, centred around 36°N and 40°N, associated with the Kuroshio Extension front (KEF) and Oyashio Extension front, respectively (see their fig. 1). During the negative phase, on the contrary, the variance of MSSTAs exhibits a broad single maximum around 40°N. However, to the north of 45°N, the variance is bigger than that during the positive phase. Note that the characteristics above are found to be robust to modest changes in the criterion we have chosen.

Yao et al. (2018) defined the SAFZ intensity through the maximum of meridional SST gradient (−∂SST/∂y) along each longitude between 145°E and 175°E. A SAFZ intensity index, $I_{int}$, was proposed, which is defined as the leading
The correlation coefficient between $I_{\text{int}}$ and MSSTA index can reach up to 0.65 (Table 1). Besides, a simple KEF index has been obtained through averaging the meridional gradient of SSTA in the KOCR (Yuan and Xiao, 2017). The correlation coefficient between KEF index and MSSTA index is as high as 0.57, passing the 95% significant test. The high correlation coefficient indicates a close link between the variability of MSSTA and the strength of ocean fronts. In the present study, we will not distinguish the SSTAs due to mesoscale oceanic eddies or fronts.

2.3 Variability of storm track

Usually, the variance or covariance of synoptic disturbances is used to represent the storm track (Chang et al., 2002). Here, we choose three metrics: the root mean square of synoptic-scale geopotential height ($\sqrt{\overline{z'}}$) at 500 hPa, the synoptic-scale meridional heat flux ($v'T'$) at 850 hPa and variance of synoptic-scale meridional wind ($v'v'$) at 300 hPa. The prime stands for 2–8-day Lanczos band-pass filter (Duchon, 1979).

We conduct empirical orthogonal function (EOF) analysis of the storm track at 500 hPa ($\sqrt{\overline{z'}}$). The first EOF mode represents the meridional shift of the storm track and explains 27.2% of the total variance, showing a dipole structure with the positive anomaly centre near (160°E, 35°N) and the negative near (160°E, 55°N). The second EOF mode represents the strengthening/weakening of storm track, with a monopole pattern accounts for 17.6% of the total variance, representing the strength variation of the storm track (not shown). However, there is a little different with the previous studies. According to the results by Lau (1988), Wettstein and Wallace (2010) and Ma and Zhang (2018), the first mode of storm track illustrates the strengthening or weakening of the storm track near its climatological centre, while the second mode demonstrates a meridional shift. The difference between our results and the previous findings is mainly caused by the period we selected. We have conducted EOF analysis for storm track at 500 hPa within the period as previous studies (from December to February), which shows a similar pattern of the leading two modes.

3 IMPACT ON THE STORM TRACK

Based on the MSSTA index defined in section 2, composite plots of eddy heat flux anomalies during the averaged positive and negative phases are shown in Figure 4. During the positive phase, the storm track anomalies exhibit a dipole pattern with a positive centre in the south and a negative centre to the north, compared to the climatology which is defined as an average during the winter time from October to February. However, the sign of the dipole pattern is opposite during the negative phase. Not only in the lower troposphere, but also at 500 or 300 hPa, the storm track shows a quite similar structure as that at 850 hPa (not shown), indicating a robust correlation between the variability of MSSTA and the meridional shift of storm track.

Moreover, the correlation coefficients between MSSTA index and first two temporal coefficients (PC1 and PC2) of EOFs are $-0.52$ and 0.26, respectively, passing the 95% significant test (Table 1). However, the position of local maxima and minima during both positive and negative phases are closer to those in EOF1 with the pattern correlation coefficient up to $-0.89$ (0.81) during the positive (negative) phase, which reveals the close relationship between the variability of MSSTA and the meridional shift of storm track.

In order to find out the cause and effect relationship between them, we computed the lead–lag correlation coefficient between the MSSTA index and the temporal coefficient of the corresponding EOF mode of the storm track. Here, the MSSTA index is calculated as above, but for all months through the years. Then the EOF analysis of the

| $I_{\text{int}}$ | KEF index | KE index | PC1 | PC2 |
|-----------------|-----------|----------|-----|-----|
| MSSTA index     | 0.65*     | 0.57*    | -0.12 | -0.52* | 0.26* |

Note: The mark * indicates 95% significance.

**FIGURE 4** Composite plots of meridional eddy heat flux for the positive phase (a) and negative phase (b) of the MSSTA index (shading: km/s). The contours represent the climatological meridional eddy heat flux (CI = 2 km/s). Statistically significant differences at 90% according to the $t$ test are stippled [Colour figure can be viewed at wileyonlinelibrary.com]
storm track has also been performed during the same period over the North Pacific (30°–65°N, 120°–240°E; not shown). The correlation coefficient reaches its minimum (−0.52) when MSSTA lead 2 months, which indicates the MSSTA variability is leading the shift of storm track. According to the results above, we will investigate the variability of MSSTA impact on the meridional shift of storm track.

4 | LARGE-SCALE ATMOSPHERIC CIRCULATION

Evidence has been emerging that mesoscale structures, such as ocean fronts, could impact not only marine atmospheric boundary layer (MABL) but also basin-scale atmospheric circulation (Nakamura et al., 2004; Sampe et al., 2010; Frankignoul et al., 2011; Taguchi et al., 2012). Nakamura et al. (2004) proposed that the difference of air–sea heat flux across the front could anchor the storm track through oceanic baroclinic adjustment mechanism. Meanwhile, the time-mean flow is coherent with the storm track (Ren et al., 2010; Li and Wettstein, 2012; Liu et al., 2014). The jet stream can significantly influence the storm track activity by altering baroclinicity (Nakamura et al., 2004).

To have some insight into the large-scale circulation, composites of SLP, geopotential height, air temperature and zonal wind are considered next (Figures 5 and 6), and the analysis indicates an equivalent barotropic structure over the North Pacific. The monthly mean composites for SLP in the positive and negative phases are shown in Figure 5a,b, respectively. The SLP exhibits a significant negative anomaly located in the vicinity of climatological Aleutian Low (referred to as the AL) in the central North Pacific during the positive phase (Figure 5a), which indicates the strengthening of the AL. Meanwhile, the positive-phase composite map of zonal wind at 850 hPa shows a dipole pattern with intensified anomaly south of the SLP anomaly (Figure 5c). The situation is opposite for the negative phase (Figure 5b,d).

The latitudinal-pressure cross section of geopotential height, air temperature and zonal wind along 160°E are shown in Figure 6 for the positive and negative phases of MSSTA index. It is notable that all three variables are consistent. During the positive phase, the negative anomaly of the geopotential height in the troposphere is accompanied by a cold anomaly that extends throughout the troposphere below 300 hPa, and the jet shifts southwards. The signs of the anomalies are all reversed during the negative phase. The dipole pattern of zonal wind from the surface to the top may indicate the meridional displacement of storm track.

The results above show a clear relationship between the variability of MSSTA and large-scale circulation. The meridional shift of storm track is accompanied by the meridional shift of the jet. In the following section, we will investigate the processes and mechanism that the MSSTAs affect the large-scale atmospheric circulation.

5 | POSSIBLE MECHANISM

5.1 | Impact of variability of MSSTAs on baroclinicity

To investigate the mechanism behind the variability of MSSTA and the meridional shift of storm track, the maximum Eady growth rate (EGR; Eady, 1949; Lindzen and Farrell, 1980) is calculated, which can be considered as a key measure of baroclinicity and is widely used for analysing

![FIGURE 5](https://example.com/figure5.png)

**FIGURE 5** Composite plots of SLP (units: hPa) and zonal wind at 850 hPa (units: m/s) for the positive phase (a, c) and negative phase (b, d) of the MSSTA index (shading). The contours in (a, b) denote the first EOF mode of SLP with the centre near the Aleutian Islands, while those in (c, d) represent the climatological of the storm track (represented by \( v^T \) at 850 hPa; CI = 2 km/s). Statistically significant differences at 90% according to the t test are stippled [Colour figure can be viewed at wileyonlinelibrary.com]
storm track dynamics (Nakamura and Yamane, 2009; Kuwano-Yoshida and Minobe, 2017; Yao et al., 2018). EGR is defined as follows:

$$\sigma_{BI} = 0.31 \frac{g f}{N} \left| \frac{\partial u}{\partial z} \right|,$$

(1)

where $N$ is the Brunt-Väisälä frequency, $g$ is the gravitational acceleration, $f$ is the Coriolis parameter and $u$ is zonal wind. According to Equation (1), the EGR is dominated by vertical wind shear and the reciprocal of the static stability.

Figure 7a,b displays the vertically integrated (from 1,000 to 700 hPa) EGR during the positive and negative phases, respectively. During the positive phase of MSSTA index, the pronounced positive anomaly is located south of the climatological storm track, which is favourable for the enhancement of baroclinicity to the south and contributes to the positive anomaly of storm track. Meanwhile, the small
negative anomaly can be found to the north of 40°N. The situation is opposite during the negative phase. Despite of the results above, we should note that EGR inside the convective MABL is sensitive to the small uncertainties in data. This problem may occur near the surface, such as 1,000 hPa. Here, we integrated EGR from 1,000 hPa to 700 hPa and the effect of undefined values did not occur in our results.

The anomaly of EGR can be explained by vertical shear in the zonal wind and static stability. During the positive (negative) phase of MSSTA index, the positive (negative) zonal wind anomaly is located to the south of climatological storm track, while the negative (positive) zonal wind anomaly is located to the north (Figures 5 and 6). That indicates the vertical shear strengthens (weakens) to the south of 40°N and weakens (strengthens) to the north, contributing to the enhancement (damping) and damping (enhancement) of baroclinicity in the south and north regions, respectively.

To manifest this feature clearly, the zonal wind anomalies at 850 hPa are shown in Figure 7c,d, overlaid on the integration of static stability which is vertically integrated from 1,000 to 700 hPa. In contrast to the zonal wind, the static stability shows negative (positive) anomalies over the western North Pacific during the positive (negative) phase of MSSTA index, which resembles the turbulent heat flux anomaly (Figure 8). According to Equation (1), the decreasing (increasing) static stability can strengthen (weak) EGR, indicating reinforcement (reduction) of baroclinicity.

During the positive phase, the static stability and zonal wind both contribute to the increase in EGR to the south of 40°N, but the two factors oppose each other to the north, which leads to a strong positive anomaly of EGR to the south and a weak negative anomaly to the north. The pattern of EGR during the negative phase is also the result of combining static stability and zonal wind, that is, the decreasing zonal wind and increasing static stability contribute positively to the strong negative anomaly of EGR to the south of 40°N, and the increasing zonal wind and increasing static stability lead to a weak positive anomaly of EGR to the north of climatological storm track.

As pointed out by Kuwano-Yoshida and Minobe (2017), the anomaly of static stability is due to the disturbance of turbulent heat flux, which is defined as the sum of sensible and latent heat fluxes (positive upwards). Figure 8a shows that the broad positive turbulent heat flux anomaly over the western North Pacific weakens the static stability in the lower troposphere and increases the EGR during the positive phase, while Figure 8b illustrates the opposite during the negative phase. The zonal mean of turbulent heat flux anomaly between 130°E and 180° is shown in Figure 8c. The maximum anomaly is found, in both phases, around 36°N where the variance of MSSTA is maximum and the KEF is located (Figure 3). Besides, the positive and negative anomalies are nearly symmetric. Also, it is easy to elucidate that the latent heat release by precipitation will contribute to the anomaly of static stability. Since there is no precipitation in NCEP/NCAR reanalysis, composites of precipitable water and specific humidity during the positive

**FIGURE 7** Composite plots of vertically integrated (from 1,000 to 700 hPa) EGR anomaly (a, b; shaded; units: 10−6 s−1) and static stability anomaly (c, d; shaded; units: K hPa−1) for the positive (a, c) and negative phase (b, d) of the MSSTA index. The contours in (a, b) represent for the climatological meridional eddy heat flux at 850 hPa (CI = 2 km/s), while zonal wind anomaly (CI = 0.5 m/s) at 850 hPa in (c, d). Statistically significant differences at 90% according to the t test are stippled [Colour figure can be viewed at wileyonlinelibrary.com]
phase and negative phase are shown in Figure 9. We can find that the precipitable water and specific humidity decrease during the positive phase, which may indicate less water vapour in the atmosphere and more precipitation. And the latent heat release by precipitation increases, contributing to the negative anomaly of static stability and thus leading to unstable atmosphere. During the negative phase, the situation is opposite.
In summary, the positive latent heat flux anomaly and precipitation diminishes the static stability and enhances the baroclinicity with the strengthening zonal wind to the south of 40°N during the positive phase. The negative surface heat flux anomaly and precipitation contribute positively to the static stability and weaken the baroclinicity with suppressed zonal wind to the south of 40°N during the negative phase.

5.2 Impact of variability of MSSTA on energy conversion

The variation of storm track is also closely related to energy conversion (Ma and Zhang, 2018; Yao et al., 2018). To further explain the impact of variability of MSSTA on the meridional shift of storm track, energy conversion processes between mean flow and eddies are investigated. Based on Cai et al. (2007), the energy conversion includes two different processes, barotropic energy conversion and baroclinic energy conversion, denoted as BT and BC, respectively. The BT and BC can be calculated as follows:

\[
BT = C_0 \left\{ \frac{1}{2} \left( v'^2 - \overline{v'^2} \right) + \left( -u'v' \right) \left( \frac{\partial v'}{\partial x} + \frac{\partial u'}{\partial y} \right) \right\},
\]

(2)

\[
BCEC_1 = -C_1 \left( \frac{\overline{P_0}}{\rho} \right)^{\frac{1}{2}} \left( -\frac{d\theta}{dp} \right)^{-1} \left( \overline{u'T} \frac{\partial T}{\partial x} + \overline{v'T} \frac{\partial T}{\partial y} \right),
\]

(3)

\[
BCEC_2 = -C_1 \left( \overline{T'f} \right),
\]

(4)

where \( C_1 = \left( \frac{\overline{P_0}}{\rho} \right)^{\frac{1}{2}}, \) and \( R, \omega, \theta, C_p, \) and \( C_v \) represent gas constant for dry air, vertical velocity, potential temperature, the specific heat capacity of dry air at the constant pressure and at the constant volume, respectively. The overbar and prime denote climatological average and transient disturbance, respectively. Equations (2)–(4) represent barotropic kinetic energy conversion (BT) between the mean flow and transient eddies, baroclinic conversion between mean available potential energy and eddy available potential energy (BC1) and baroclinic conversion between eddy available potential energy and transient kinetic (BC2), respectively. All three terms are computed at 850 hPa in the present study.

The magnitude of barotropic energy conversion (BT; not shown) is much smaller than the baroclinic energy conversion, both BC1 and BC2. Hence, we will focus on the baroclinic energy conversion. Figure 10 shows BC1 and BC2 during the positive and negative phases of MSSTA index. During the positive phase, BC1 and BC2 display a similar dipole structure over the western North Pacific with a significant positive anomaly along the southern side of the climatological storm track. That indicates that there is more BC1 during the positive phase than that in normal period, contributing to increase BC2 south of the 40°N during the positive phase. More transient eddy kinetic energy to the south strengthens storm track there, exhibiting a southwards shift (Figure 4a). On the contrary, the baroclinic energy conversion decreases south of 40°N notably during the negative phase.

**FIGURE 10** Composite plots of baroclinic energy conversion, BC1 (a, b; W/m²) and BC2 (c, d; W/m²) for the positive phase (a, c) and negative phase (b, d) of the MSSTA index. The contours represent the climatological meridional eddy heat flux at 850 hPa (CI = 2 km/s). Statistically significant differences at 90% according to the t test are stippled [Colour figure can be viewed at wileyonlinelibrary.com]
phase of MSSTA index, leading to the weakening of the storm track (Figure 4b).

The results in this section further investigated the meridional shift of the storm track. The baroclinic energy conversion, BC1 and BC2, explains the significant anomaly of storm track to the south of 40°N. However, the less pronounced anomalies to the north over the whole period could not be explained by the above energy arguments. Perhaps another mechanism can bridge the gap.

5.3 | Synoptic-scale response

The main feature of the storm track response to the variability of MSSTA is a meridional shift shown in section 3. We explore the synoptic-scale response to the south and north of 40°N. The in-phase coherent pattern in both phases indicates the PAM mechanism, which is dominant south of the storm track. Figure 11a,c shows the synoptic SLP anomalies, omega and divergence averaged over south of the climatological storm track (30°–35°N) for the positive phase. The red line denotes the boundary layer height. We can see that there is a negative anomaly of divergence near 170°E, contributing negative value of omega (upwards) penetrating to the free atmosphere over the synoptic SLP trough at 170°E (Figure 11a,c). At the same time, there is a positive anomaly of divergence located above a ridge at 140°E along with positive (downwards) vertical velocity anomaly. But, the synoptic divergence anomalies induced by SLP anomalies are not only confined in the MABL; they could reach up to 700 hPa in Figure 11a. Furthermore, the synoptic divergence anomalies are significantly stronger in the upper troposphere near 250 hPa. The divergence and vertical velocity anomalies clearly show a secondary circulation from the MABL to the upper troposphere. For the negative phase, the relationship is the same, but with a smaller magnitude for the vertical velocity west of the dateline.

By comparing Figure 11a with Figure 11b, it is evident that the synoptic disturbance could transport further downstream during the positive phase than during the negative phase. This contrasts with O’Reilly and Czaja (2015), who showed the baroclinic Rossby wave could develop rapidly over the western North Pacific and extend less far downstream during the positive phase of KEF, which indicates a stronger SST front along the KE. The difference may be due to the methods that produced MSSTA index and KEF index. The MSSTA index not only represents the intensity of ocean fronts but also the variability of ocean eddies. Therefore, the atmosphere response in the present study should be regarded as the combined effect of the two kinds of ocean mesoscale structures: ocean front and mesoscale eddy.

Similarly, Figure 12 shows the composite plots of synoptic SLP, divergence and vertical velocity, averaged north of the climatological storm track (45°–55°N). Compared with Figure 11, the vertical motion and divergence are weak between 120°E and 180° during the positive phase (Figure 12a). On the contrary, during the negative phase
(Figure 12b), there are a trough and a ridge near 150°E and 170°E, respectively. However, the convergence does not coincide with the SLP trough near 150°E, but over 160°E where the SLP anomaly is nearly zero. At the same time, the significant vertical motion is induced because of the convergence below the MABL, and penetrates into the free atmosphere, up to 200 hPa where divergence is strong. All these consist of the secondary circulation north of the climatological storm track. Figure 3d shows that the variance of MSSTA is stronger to the north of 45°N during the negative phase than during the positive phase, which may be the reason for the significant upwards motion around 160°E.

Figure 13 displays the vertical profiles of eddy momentum flux and eddy heat flux along 160°E. Both eddy momentum flux and eddy heat flux anomalies feature a meridional dipole structure. During the positive phase, the positive anomaly of eddy momentum flux is located near 30°–40°N with the centre near 500 hPa; and the negative anomaly exists between 45°N and 60°N with the centre located higher. As for the negative phase, the anomalies exhibit broadly from 1,000 to 200 hPa, with a positive anomaly to the north of 40°N and a negative to the south. Comparing Figure 13a with Figure 13b, the anomalies are much stronger in the negative phase, especially to the north of 40°N. The positive anomaly to the north of 40°N indicates upwards eddy momentum flux, which contributes to the strengthening of storm track during the negative phase. The significant upwards motion near 160°E by the VMM may be one of the reasons for the positive anomaly of eddy momentum flux. On the other hand, the upstream transient disturbances may be transported to the western North Pacific through the waveguide as indicated by Chang and Yu (1999); however, it is beyond the scope of this paper.

Figures 13c,d illustrates the anomalies of eddy heat flux during the positive and negative phases, respectively. Compared with the eddy momentum flux, the anomalies of eddy heat flux exhibit a similar dipole pattern during both phases, but the anomaly centres are at a lower level. Besides, the anomalies to the south of 40°N, which are located around 35°N, are more intense than those to the north. The anomaly centres of turbulent heat flux are spatially coherent with the anomalies of eddy heat flux to the south of 40°N, which could be an explanation for the distribution of anomalies of eddy heat flux. More (less) heat out of the ocean is penetrated into the atmosphere during the positive (negative) phase, enhancing (weakening) the vertical eddy heat flux to the south of the climatological storm track in the lower troposphere.

Overall, the results show synoptic SLP, divergence and vertical velocity are spatially in phase to the south of the climatological storm track during both phases. However, to the northern side of the climatological storm track, things are different. Vertical velocity is out of phase with the SLP anomaly during the negative phase while the vertical velocity and divergence are very weak during the positive phase. The strong vertical motion to the north of 40°N during the
negative may force vigorous eddy momentum flux anomaly, modulating the fluctuation of storm track along the northern side of its climatological position. These results indicate there may be different mechanisms to the north and south of climatological storm track. At this point, we can only speculate both the VMM and PAM are at play. However, for the relative low resolution of NCEP/NCAR reanalysis, we may not isolate the mechanisms.

6 CONCLUDING REMARKS AND DISCUSSION

In the present study, the effect of the variability of MSSTA in the KOCR on the meridional shift of the North Pacific storm track is investigated. An index of the variability of MSSTA is proposed using a high-resolution SST data set. In the positive phase of the MSSTA index, the variance of MSSTAs are stronger and features double peaks within the KOCR, whereas in the negative phase, the variance of MSSTA is relatively weak. Using the MSSTA index, the response of storm track and the associated mechanisms are examined. During the positive phase of the MSSTA index, the storm track shifts southwards, exhibiting a significant dipole structure with a negative anomaly to the north of its climatological position and a positive anomaly to the south. The situation is opposite during the negative phase of the MSSTA index.

To understand the meridional shift of storm track associated with MSSTAs, the surface air–sea heat fluxes, baroclinicity, energy conversion processes and synoptic response are examined. It is revealed that the strengthening (weakening) turbulent heat flux and more (less) latent heat release by precipitation during the positive (negative) phase of MSSTA index, transfer more (less) heat into the atmosphere along the southern side of the climatological storm track, which leads to the weakening (strengthening) of static stability and hence contributes to the intensification (damping) of baroclinicity to the south. Meanwhile, the large-scale circulation exhibits a southwards (northwards) shifted jet during the positive (negative) phase of MSSTA index, contributing to the enhancement (decrease) of baroclinicity to the south of 40°N. The fluctuation of baroclinicity is tightly related to the development of the storm track. The enhancing (decreasing) baroclinicity contributes to the strengthening (weakening) of the storm track through baroclinic energy conversion, exhibiting a significant positive (negative) anomaly (Figure 4) to the south of 40°N during the positive (negative) phase of MSSTA index.

Previous studies mostly focused on the influence of ocean fronts on the Pacific storm track (Joyce et al., 2009;
Frankignoul et al., 2011; O’Reilly and Czaja, 2015; Révelard et al., 2016; Yao et al., 2018). What is the relationship between the ocean fronts and the mesoscale oceanic eddies? Here, we try to discuss this question briefly. Qiu et al. (2014) proposed a KE index to represent the KE state, which indicates that the KE jet is strong and has a steady and northerly path in its stable state, and has reversed properties during the unstable state. The correlation coefficient between MSSTA index and KE index is only −0.12 (Table 1), which indicates little consistency between the KE state and the variability of MSSTA. The poor correlation between the MSSTA index and KE index may be due to the domain selected for calculating the indexes. The KE index is based on the momentum fields and reflects the axis of the KE, and the region used for defining it is confined within a narrow band. Nonaka et al. (2006) indicated that the KE front is the strongest in depth between 200 and 600 m. The variability of SSTA over the larger region of the KOCR may be not correlated with the KE. Despite of little consistency, the composite plots of SST during the positive and negative phase of MSSTA index do partially indicate the meridional shift of the KE (Figure 14), with colder SST during the positive phase of MSSTA index indicating a southerly KE path and an unstable state. When the KE is in an unstable state, more eddies detach from the KE front and move northwards (Itoh and Yasuda, 2010), which may contribute to larger variability of SSTA during the positive phase of MSSTA index.

Furthermore, the MSSTA index shows a significant positive correlation with the SAFZ intensity index proposed by Yao et al. (2018). The SAFZ is associated with the Oyashio Extension and is identified by SST gradient. The regions used for defining the two indexes are similar. The SST front may contribute to the variability of MSSTA. But the response of the storm track is not identical, and the oceanic mesoscale eddies can be responsible for this gap. Overall, as filtered by the boxcar filter above, the MSSTAs could not avoid the impact of ocean fronts. In addition, the response of atmosphere should be regarded as the result of the combined effect of mesoscale eddies and fronts.

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