

Effects of Tropical Sea Surface Temperature Variability on Northern Hemisphere Tropical Cyclone Genesis

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ABSTRACT: This study quantifies the contributions of tropical sea surface temperature (SST) variations during the boreal warm season to the interannual-to-decadal variability in tropical cyclone genesis frequency (TCGF) over the Northern Hemisphere ocean basins. The first seven leading modes of tropical SST variability are found to affect basinwide TCGF in one or more basins, and are related to canonical El Niño–Southern Oscillation (ENSO), global warming (GW), the Pacific meridional mode (PMM), Atlantic multidecadal oscillation (AMO), Pacific decadal oscillation (PDO), and the Atlantic meridional mode (AMM). These modes account for approximately 58%, 50%, and 56% of the variance in basinwide TCGF during 1969–2018 over the North Atlantic (NA), northeast Pacific (NEP), and northwest Pacific (NWP) Oceans, respectively. The SST effect is weak on TCGF variability in the north Indian Ocean. The SST modes dominating TCGF variability differ among the basins: ENSO, the AMO, AMM, and GW are dominant for the NA; ENSO and the AMO for the NEP; and the PMM, interannual AMO, and GW for the NWP. A specific mode may have opposite effects on TCGF in different basins, particularly between the NA and NEP. Sliding-window multiple linear regression analyses show that the SST effects on basinwide TCGF are stable in time in the NA and NWP, but have strengthened since the 1990s in the NEP. The SST effects on local TC genesis and occurrence frequency are also explored, and the underlying physical mechanisms are examined by diagnosing a genesis potential index and its components.

KEYWORDS: Atmosphere–ocean interaction; Tropical cyclones; Climate variability; Sea surface temperature

1. Introduction

Around 70% of global tropical cyclones (TCs) are in the Northern Hemisphere [including the North Atlantic (NA), northeast Pacific (NEP), and northwest Pacific (NWP) Oceans and the north Indian Ocean (NIO)]. Their genesis is strongly modulated by the large-scale atmospheric environment in which the storms are embedded and by the surface temperature of the underlying ocean from which the storms extract energy for development (e.g., Gray 1968; Zehr 1992; Emanuel 2003, 2018; Knutson et al. 2008, 2010; Tory and Frank 2010; Vecchi et al. 2011; Nolan and McGauley 2012). The large-scale atmospheric environment is also shaped by the surface temperature of the ocean, particularly in the tropics (e.g., Palmer and Mansfield 1984; Lau 1985; Kucharski et al. 2008; Xie et al. 2016; Wang 2019). Thus, a good understanding and an accurate prediction of tropical sea surface temperatures (SSTs) can improve the prediction of seasonal TC genesis frequency

(TCGF). Indeed, year-to-year variations in TCGF have been linked to various modes of SST variability, such as El Niño–Southern Oscillation (ENSO), the Atlantic meridional mode (AMM), Atlantic multidecadal oscillation (AMO), Pacific decadal oscillation (PDO), the Pacific meridional mode (PMM), and the Indian Ocean basin (IOB) mode (e.g., Camargo et al. 2010).

Climate modes that have been identified as influencing NA TCGF include ENSO, the AMM, and AMO. SST anomalies in the equatorial Pacific associated with El Niño displace the atmospheric convection eastward in the tropical Pacific, which weakens the Walker circulation and strengthens westerlies in the upper troposphere over the tropical NA and Caribbean Sea. The enhanced westerly increases vertical wind shear in these regions and thereby suppresses TC genesis in the NA (e.g., Gray 1984; Goldenberg and Shapiro 1996; Landsea et al. 1999; Ayyer and Thorncroft 2006; Klotzbach 2011). Meanwhile, El Niño tends to stabilize the air column in the tropical NA by warming up the upper troposphere (Tang and Neelin 2004; Klotzbach 2011) and reduce tropospheric humidity (Camargo et al. 2007), both of which also discourage TC genesis in this region. In contrast, NA TC genesis is active during a positive phase of the AMM, which features warmer-than-normal SSTs over the tropical NA and a northward shift of the intertropical convergence zone (Xie and Carton 2004; Chiang and Vimont 2004). The anomalously warm SSTs in

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the tropical NA tend to reduce sea level pressure and enhance deep convection, resulting in an anomalous atmospheric circulation that decreases vertical wind shear and increases low-level vorticity. All of these favor TC genesis in the NA (Xie et al. 2005; Kossin and Vimont 2007; Vimont and Kossin 2007). The AMO has nearly the same effects as the AMM but is more prominent on multidecadal time scales and has contributed to the above-normal TC activity since the mid-1990s when it switched to a positive phase (Landsea et al. 1999; Goldenberg et al. 2001; Klotzbach 2011).

In the NEP, TC genesis is modulated by SSTs associated with climate modes in both the Pacific and Atlantic. ENSO tends to influence TC genesis in the western NEP via changes in local SSTs, convective activity and relative humidity, with higher TC genesis during El Niño years (Watterson et al. 1995; Clark and Chu 2002; Camargo et al. 2008; Klotzbach and Blake 2013; Jien et al. 2015). The ENSO effect is weak on TCGF in the eastern NEP (Collins 2010; Caron et al. 2015; Jien et al. 2015). As a result, in the NEP El Niño shifts the mean TC genesis location westward (Irwin and Davis 1999) with only a modest difference in basinwide TCGF between El Niño and non-El Niño years (Whitney and Hobgood 1997). Meanwhile, SST anomalies associated with the AMO and in the NA subtropical gyre can modulate NEP TC genesis on both interannual and decadal time scales, with more TCs forming in the NEP when NA SSTs are below normal. These Atlantic factors operate by changing upper-level easterly winds and thus vertical wind shear in the NEP (Camargo et al. 2007, 2008; Wang and Lee 2009; Caron et al. 2015; Zhang and Wang 2015). In short, the effects of ENSO and NA SSTs on the NEP TC genesis are opposite to those on NA TC genesis, which can largely explain the observed negative correlation in TC counts between the two basins (Elsner and Kara 1999; Wang and Lee 2009; Collins 2010; Caron et al. 2015). In addition, the PMM may also exert a strong influence on NEP TC genesis and occurrence frequency, particularly in the western portion (Collins et al. 2016; Murakami et al. 2017a; Wood et al. 2019).

The factors governing TC genesis in the NWP mainly include ENSO, the PMM, and SST anomalies in the Indian and Atlantic Oceans. ENSO tends to generate more TCs in the southeastern quadrant of the NWP and fewer TCs in the northwestern quadrant during its positive phase, without significantly altering basinwide TC counts (e.g., Chan 1985; Wang and Chan 2002; Choi et al. 2015). Such a displacement in genesis location is associated with increased low-level vorticity and reduced vertical wind shear in the southeastern NWP (e.g., Wang and Chan 2002; Clark and Chu 2002), tied to an eastward expansion of the monsoon trough, and with decreased midlevel humidity near the Asian continent (Camargo et al. 2007). Since the 1990s, a different flavor of El Niño has occurred frequently with the warm anomaly peaking in the central equatorial Pacific (e.g., Yeh et al. 2009; Yu et al. 2012). This central-Pacific (CP) El Niño (also called El Niño Modoki; Ashok et al. 2007; Kao and Yu 2009) favors higher basinwide TCGF (e.g., Chen and Tam 2010; Kim et al. 2011; Mei et al. 2015; Wu et al. 2018; Zhao and Wang 2019) by exciting conducive atmospheric condition (e.g., cyclonic circulation anomaly in the lower troposphere) over much of the basin.

More recently, the PMM, which features a meridional dipole pattern of SST anomalies and sometimes may promote the initiation of CP or canonical El Niño (e.g., Larson and Kirtman 2014; Capotondi and Sardeshmukh 2015; Amaya 2019), has been linked to the variations in NWP TCGF: a positive PMM encourages more TC genesis primarily by changes in dynamical factors (e.g., decreased vertical wind shear; Zhang et al. 2016; C. Liu et al. 2019). SST anomalies outside the Pacific also exert influences on NWP TCGF (Wu et al. 2020). Anomalous SST warming in the tropical Indian Ocean during post-El Niño summers (Xie et al. 2009) can significantly suppress NWP TCs by triggering an anomalous anticyclone over the NWP (e.g., Du et al. 2011; Zhan et al. 2011; Tao et al. 2012; Ha et al. 2015). Above-normal SSTs in the tropical NA may also discourage NWP TC genesis via their effects on SSTs of the Indian Ocean and subtropical North Pacific and on the Walker circulation (e.g., Ham et al. 2013; Huo et al. 2015; Yu et al. 2016; Gao et al. 2018; Zhang et al. 2018).

ENSO is the only climate mode that has been linked to the interannual variations in TCGF over the NIO. In the Bay of Bengal, fewer TCs form in the months of May and November under El Niño conditions, in response to reduced low-level relative vorticity, increased vertical wind shear, and reduced relative humidity; the effect is insignificant in other months (Singh et al. 2000; Felton et al. 2013). The effect of ENSO on TCGF in the Arabian Sea is very weak (Singh et al. 2000). Consistently, the genesis potential indices (GPIs), reanalysis products, and dynamical models (either coupled or atmosphere-only) cannot reproduce the observed year-to-year TCGF variations over the NIO, although some of them are able to capture the climatological seasonal cycle of TC genesis (Camargo et al. 2007; Zhao et al. 2009; Camp et al. 2015). These results suggest that either the NIO TCGF is unpredictable or other factors (including those related to the underlying SSTs) still await identification (e.g., Murakami et al. 2017b).

As stated above, several tropical SST factors have been identified as affecting TCGF variability, and previous studies tend to focus on a small subset of these factors. It is unclear whether these SST factors are independent. Moreover, their relative contributions to TCGF variations remain to be quantified and the underlying physical mechanisms need to be explored. To address these issues and quantify the integrated effects of the tropical SSTs on TC activity, we take the tropical oceans as a whole, and first obtain the leading modes that effectively and thoroughly represent tropical SST variations via a systematic and objective decomposition. Specifically, we extract the leading modes in the tropical SSTs during the boreal warm season (i.e., June–November) using an empirical orthogonal function (EOF) analysis (section 3). We then quantify and compare the individual and combined effects of these distinct SST modes on TC genesis and occurrence frequency in the four basins of the Northern Hemisphere (section 4a). Insights into how tropical SSTs modulate TC genesis in different basins are gained through a diagnosis and decomposition of a GPI (section 4b). Last, we examine the potential decadal changes in these tropical SST effects (section 4c).

2. Data and methods

a. Data

Monthly SSTs of the boreal warm season (i.e., June–November, the period when nearly 90% of Northern Hemisphere TCs form) during 1951–2018 from the Hadley Centre SST dataset (Rayner et al. 2003) are used to study the interannual-to-decadal variations in tropical SSTs during the Northern Hemisphere TC peak season. Monthly atmospheric variables during June–November from the Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) are adopted to connect variations in TCGF to those in tropical SSTs and understand the underlying physical mechanisms. Using the Extended Reconstructed Sea Surface Temperature version 5 (ERSSTv5; Huang et al. 2017) dataset and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR; Kalnay et al. 1996) Reanalysis 1 dataset produces results very similar to those obtained using the Hadley Centre SST and JRA-55 datasets.

TC data for the NA and NEP (east of 180° longitude) basins are from the National Hurricane Center best-track dataset (Landsea and Franklin 2013), and for the NIO are from the Joint Typhoon Warning Center (JTWC) best-track dataset (Chu et al. 2002). For the NWP basin (west of 180° longitude), because of the discrepancies in TC frequency among different data sources we use three best-track datasets respectively from the Shanghai Typhoon Institute of the China Meteorological Administration (CMA), Japan Meteorological Agency (JMA), and JTWC (e.g., Ren et al. 2011; Schreck et al. 2014; Barcikowska et al. 2017). All these best-track datasets provide the location and intensity of TCs at 6-h intervals. TCs with at least tropical storm intensity during 1951–2018 are considered in this study, but since satellite measurements became available after the late 1960s, only those over the most recent 50 years (i.e., 1969–2018) are used in the regression analysis (section 4) given the higher quality of the data (e.g., Chang and Guo 2007; Landsea et al. 2010; Vecchi and Knutson 2011). For the NIO, only TC data since 1977 are used because of the inconsistencies between JTWC archive data and the data published in JTWC's annual tropical cyclone reports (Chu et al. 2002). It is worth noting that the short-duration tropical storms increased since 2000 in North Atlantic while medium- to long-lived storms have increased little, which has been attributed to changes in observing capabilities (Landsea et al. 2010; Villarini et al. 2011). To account for these biases in TC measurements, we consider only TCs with a duration of at least two days.

b. Methods

An EOF analysis is employed to extract the leading modes of interannual-to-decadal variability in boreal warm-season-averaged SSTs over the tropics (30°S–30°N) during 1951–2018. Multiple linear regression analysis is used to build statistical models linking TCGF in individual basins to the principal components (PCs) of the leading SST modes. Simple linear regression and correlation analyses are applied to identify the atmospheric conditions that physically connect tropical SST variations with local TC genesis and occurrence

frequency, with the significance level of the results estimated by the standard two-tailed Student's *t* test.

TC genesis frequency is calculated as the total number of TCs forming in each basin or each 8° × 8° grid on a yearly basis. TC occurrence frequency (also called track density) is calculated as the total number of TC days each basin or each 8° × 8° grid experiences on a yearly basis. The large grid (i.e., 8° × 8°) is used in order to reduce the noise level and using a smaller grid (e.g., 5° × 5° or 6° × 6°) gives similar results. A GPI, denoting the favorability of the large-scale environment in which TCs develop, is calculated following Emanuel (2010) as

$$\text{GPI} = \frac{a|\eta|^3[\max(V_{\text{PI}} - 35, 0)^2]}{\chi^{4/3}(25 + V_{\text{sh}})^4}, \quad (1)$$

where *a* is a constant and is set to be 10¹⁶ in this study, η is the 850-hPa absolute vorticity, V_{PI} is the TC potential intensity, χ is the 600-hPa entropy deficit, and V_{sh} is the magnitude of the 250–850-hPa wind shear vector (see also Korty et al. 2012; Tang and Emanuel 2012).

3. Tropical SST modes

The first seven modes (M1–M7) extracted via the EOF analysis totally account for 79.9% of the variance in tropical SST anomalies, with each explaining 39.6%, 20.2%, 7.0%, 3.8%, 3.4%, 3.2%, and 2.7%, respectively. These tropical SST modes can be linked to well-known modes of climate variability, although M4 to M7 may not be well separated based on North's rule of thumb (North et al. 1982) presumably because of 1) the short study period and 2) the locality of these SST modes; indeed, Messié and Chavez (2011) were able to obtain modes that are similar but significant with 100-yr monthly data. Only one mode is tied to the secular global warming trend, whereas three modes are dominated by variations in the Pacific and the other three by variations in the Atlantic. Collectively, these SST modes well represent the signal in tropical SST variations. Before quantifying their combined effects on the Northern Hemisphere TC activity, we first describe these modes individually. The left and right panels of Fig. 1 show their spatial patterns and PCs (PC1–PC7), respectively.

Mode M1 explains 39.6% of the tropical SST variance and features the classical ENSO pattern with the strongest variability in the central and eastern equatorial Pacific (Fig. 1a). Its PC (i.e., PC1) is highly correlated with Niño indices (e.g., correlation coefficient $r = 0.93$ for the Niño-3 index during 1951–2018; Fig. 1b) and captures strong El Niño events (e.g., in 1997 and 2015). The spatial pattern of this mode resembles that of the leading mode extracted from global detrended SST anomalies of 1910–2009 in Messié and Chavez (2011), indicating the robustness of this mode. Meanwhile, an Indian Ocean dipole (IOD)-like pattern exists over the Indian Ocean (Fig. 1a) and PC1 is significantly correlated with the IOD index ($r = 0.57$; Fig. 1b), consistent with previous studies suggesting that ENSO and the IOD are coherently linked (e.g., Ju et al. 2004; Y. Liu et al. 2019).

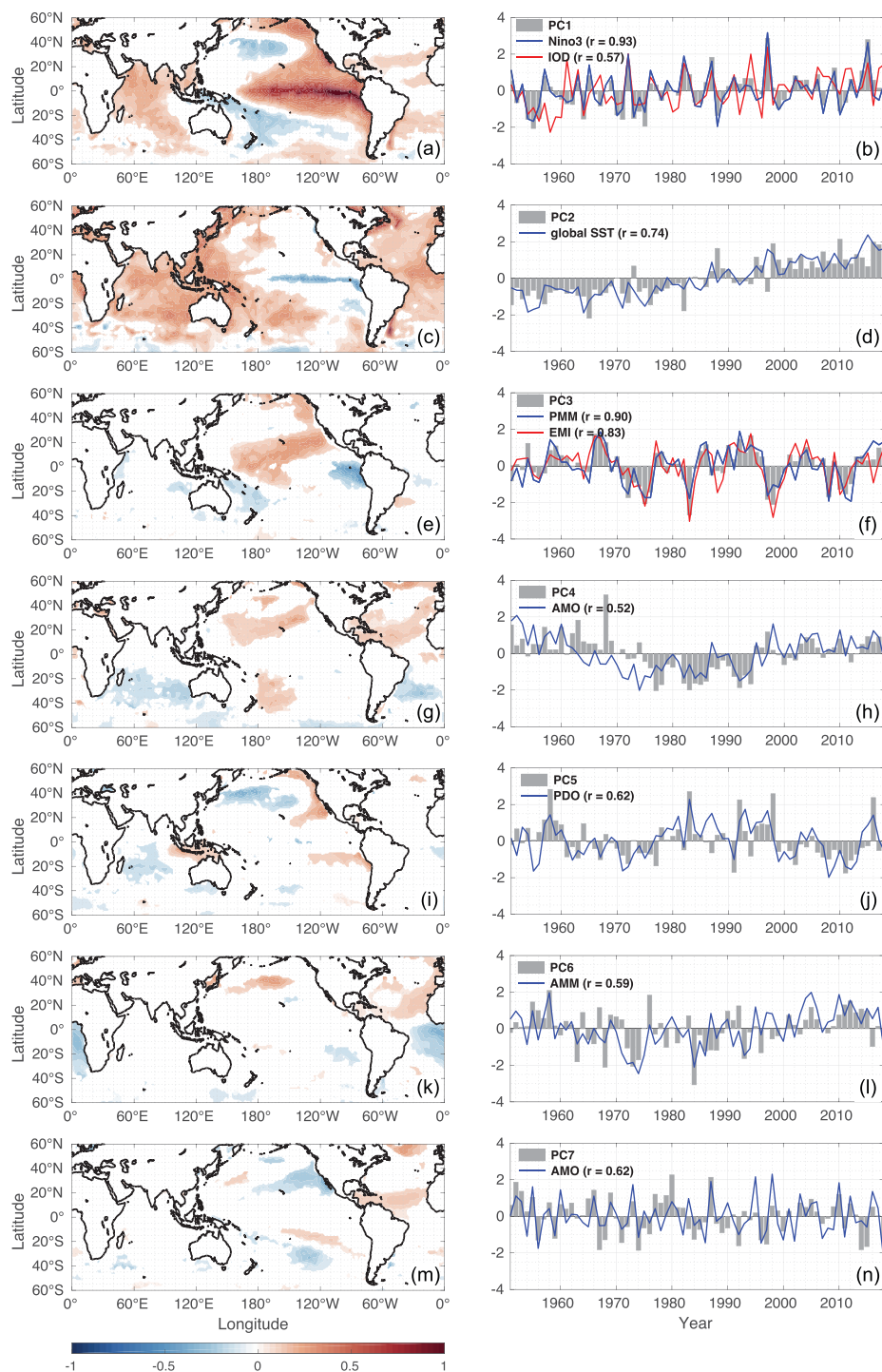


FIG. 1. (left) The boreal warm-season (i.e., June–November) SST anomalies regressed against the principal components (PCs) of the first seven leading modes (M1–M7) of tropical SSTs during 1951–2018: (a) PC1, (c) PC2, (e) PC3, (g) PC4, (i) PC5, (k) PC6, and (m) PC7. Only values significant at the 0.05 level are shown (shading). (right) The normalized PCs (gray bars) of M1–M7, also showing normalized climate indices: (b) Niño-3 index (blue) and the IOD index (red), (d) global-mean SST (blue), (f) the PMM index (blue) and EMI (red), (h) the detrended AMO index (blue), (j) the PDO index of April–July (blue), (l) the AMM index (blue), and (n) the 8-yr high-pass-filtered AMO index (blue).

Mode M2 features a pattern of global warming with strong warming occurring during the past seven decades in the Atlantic Ocean, Indian Ocean, and western portion of the Pacific Ocean. Weak cooling occurs in the equatorial eastern Pacific Ocean and Southern Ocean (Fig. 1c). PC2 shows a strong upward trend and is highly correlated with the global mean SST during 1951–2018 ($r = 0.74$; Fig. 1d). It is worth noting that, when individual basins are considered, PC2 is significantly correlated with the Atlantic Niño index ($r = 0.62$) and IOB index ($r = 0.71$) during 1951–2018, implying the covariations of SSTs in the tropical Atlantic and Indian Oceans. This mode shares similarities with a combination of EOF modes 1, 2, and 3 in Dong and Zhou (2014) and the second mode of tropical Pacific SST variability shown in Jiang and Zhu (2018). We further separated PC2 into interannual and decadal components and then regressed them against global SST anomalies (not shown). It is found that the interannual component is associated with La Niña and significantly correlated with PC1 ($r = -0.54$). The regressed SST pattern of the low-frequency component resembles that of Dong and Zhou (2014). It is unclear why part of ENSO variability is embedded in this global warming mode.

Mode M3 resembles the PMM, which is the second leading mode of coupled ocean–atmosphere variability in the Pacific and characterized by out-of-phase SST variations over the north tropical Pacific and eastern equatorial Pacific (Fig. 1e; Chiang and Vimont 2004; Amaya 2019). The correlation coefficient between PC3 and the PMM index defined by Chiang and Vimont (2004) is 0.90 for the period of 1951–2018 (Fig. 1f). During a positive phase, the warm anomaly extends to the central equatorial Pacific and cold anomaly exists over eastern and western (though insignificant for the latter) equatorial Pacific, thus this mode also shares some similarities with the CP ENSO (Ashok et al. 2007; Kao and Yu 2009); the correlation coefficient between PC3 and the El Niño Modoki index (EMI) is 0.83 during 1951–2018 (Fig. 1f).

Mode M4 exhibits some typical characteristics of the AMO and features anomalous warming in the tropical and high-latitude North Atlantic and in the subtropical North Pacific, and a cold anomaly over the subtropical South Atlantic and Indian Ocean (Fig. 1g). PC4 primarily displays low-frequency variations, and its correlation coefficient with the detrended AMO index is 0.52, with the two time series differing mainly during the 1960s and early 1970s (Fig. 1h). When both time series are filtered using an 8-yr low-pass filter, their correlation coefficient rises to 0.65 (not shown).

Mode M5 is characterized primarily by a dipole pattern of SST variations over the midlatitude western and eastern North Pacific (Fig. 1i). This mode is very similar to the third mode of the global SST variability depicted in Messié and Chavez (2011), and both differ from the conventional PDO, which also exhibits strong SST variability in the equatorial Pacific. The correlation coefficient between PC5 and the PDO index of June–November and of April–July during 1951–2018 is 0.46 and 0.62 (Fig. 1j), respectively. An east–west dipole pattern also exists over the low-latitude south Indian Ocean (Fig. 1i).

Mode M6 resembles the AMM, with SST changes in the tropical NA opposite to those in the tropical South Atlantic (Fig. 1k). The correlation coefficient between PC6 and the AMM index defined based on the coupled ocean–atmosphere variability (Chiang and Vimont 2004) is 0.59 during 1951–2018 (Fig. 1l). The spatial pattern of this mode is similar to the opposite phase of the sixth mode in Messié and Chavez (2011), where it is described as the Atlantic Niño; the correlation coefficient between PC6 and the Atlantic Niño index is -0.59 (not shown).

Mode M7 features anomalous SST warming in the NA (Fig. 1m), similar to the pattern of M4 but primarily on interannual time scales; the correlation coefficient between PC7 and the interannual component of the AMO index is 0.62 during 1951–2018 (Fig. 1n). High loadings are also over the subtropical NEP, where significant cooling is observed (Fig. 1m). This cooling can be linked to the warming in the tropical NA via interbasin ocean–atmosphere interactions (Ham et al. 2013): SST warming over the tropical NA generates an anomalously cyclonic circulation in the lower troposphere over the subtropical NEP as a Gill-type Rossby-wave response, and the northerlies on the west flank of the anomalous cyclone produce SST cooling via both enhanced wind speed (and thus enhanced evaporation) and cold advection from higher latitudes.

The above SST modes are statistically independent over the 68-yr period. Because of the limited realizations, especially for interdecadal variations and unfolding global warming, uncertainty remains as to the detailed spatial structure of M2. Various techniques have been attempted to ascertain the global warming pattern (e.g., Tokinaga et al. 2012; Chen and Wallace 2016). These caveats aside, the EOF is an effective and efficient method to decompose SST variations into a small number of modes. Using the ERSSTv5 dataset produces very similar results (see Fig. S1 in the online supplemental material), confirming the robustness of the identified tropical SST modes. We note that the PCs of some of the extracted modes (particularly modes 4–7) are not very highly correlated with the indices of the related climate modes, but for convenience of reference, in the following sections we refer to the first seven leading modes as ENSO, the global warming pattern, PMM, AMO, PDO, AMM, and interannual AMO, respectively.

4. Tropical SST effects on Northern Hemisphere TC genesis and occurrence frequency

a. Effects on basinwide TC genesis and occurrence frequency

As discussed in the introduction, tropical SSTs exert strong influences on TCGF. Next, we will quantify the combined effects of the leading tropical SST modes on basinwide TCGF in individual basins via a multiple linear regression analysis. Because TC data are more reliable after the advent of the satellite era, the regression analysis is applied to the most recent 50 years, that is, 1969–2018 (1977–2018 for the NIO; see section 2a for a discussion on NIO TC data). The multiple

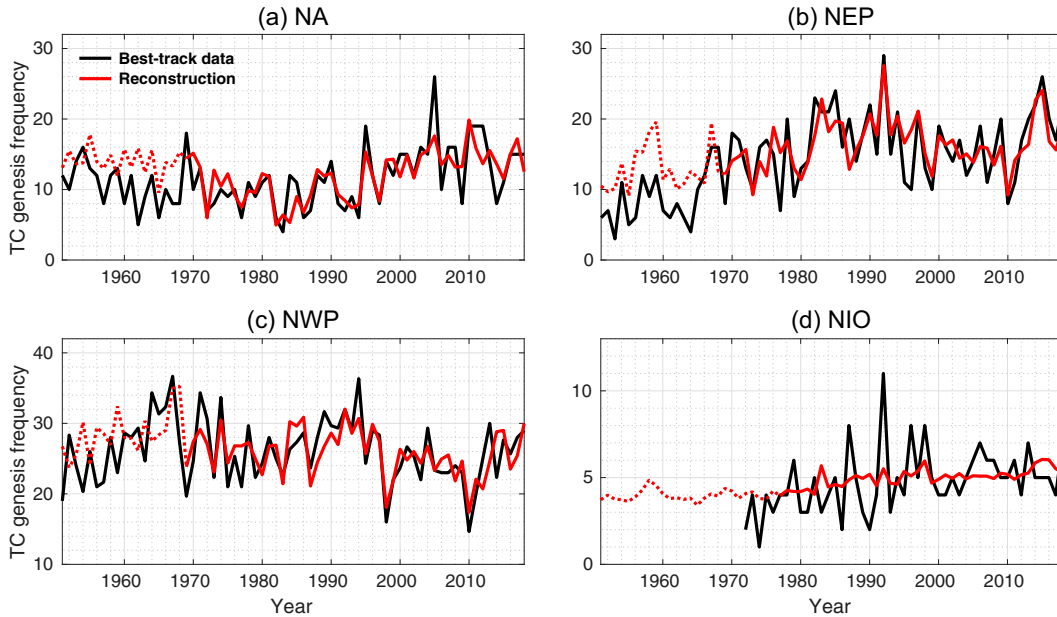


FIG. 2. The observed (black) and reconstructed (red) TCGF during 1951–2018 for the (a) NA, (b) NEP, (c) NWP, and (d) NIO. The reconstruction is based on the multiple linear regression of TCGF and PCs of the seven leading SST modes (gray bars in Fig. 1) during 1969–2018 (1977–2018 for the NIO).

linear regression analysis also allows us to quantify the relative importance of different tropical SST modes in TCGF over different basins.

Black curves in Fig. 2 show the temporal evolution of annual TCGF in the best-track data during 1951–2018 over the four basins of the Northern Hemisphere. In the NIO, TCGF varies substantially on interannual time scales (the coefficient of variation is 0.35); in the other three basins, TCGF exhibits strong variations on both interannual and decadal time scales. The TCGF reconstructed from the multiple linear regression analysis well reproduces the observations in the latter three basins, with the correlation coefficients around 0.77, 0.73, and 0.75 for the NA, NEP, and NWP during 1969–2018 (or 59%, 53%, and 56% in terms of explained variance; Figs. 2a–c), respectively. We note that if we assume the roles of the seven leading modes do not change with time (discussed in section 4c), the dotted red curves for the period before 1969 indicate that some TCs may be missing in the pre-satellite era in these three basins, which is consistent with previous studies using various distinct approaches (e.g., Chang and Guo 2007; Vecchi and Knutson 2008, 2011; Mei et al. 2019). A quantitative assessment of the “missed” TCs based on the statistical reconstruction using tropical SST modes will be explored in the near future and presented in a separate paper. In the NIO, however, the reconstruction does not perform well in capturing the year-to-year variations in observed TCGF (Fig. 2d). The lack of skill over the NIO suggests a rather modest role of tropical SSTs in modulating TCGF over this basin and echoes the findings of previous studies (e.g., Camargo et al. 2007; Zhao et al. 2009; Camp et al. 2015; Murakami et al. 2017b). From now on, we will focus our analysis on the three basins of the North Atlantic and North Pacific.

Figure 3a shows the regression coefficients and the statistical significance for the NA, NEP, and NWP. The AMO, ENSO, the AMM, and the global warming pattern (including SST warming in the NA) are important for NA TCGF; the order of these four modes reflects their relative contributions (from high to low) to the interannual-to-decadal variability in NA TCGF (figure not shown). Specifically, a positive AMO, a negative ENSO, a positive AMM, and global warming pattern of the past five decades favor NA TC genesis.

We have reconstructed the SST pattern conducive to NA TC genesis using the spatial patterns of the seven leading SST modes (M1–M7) as follows:

$$\text{SST}_{\text{reconstruction}} = [(T - 1)\sigma_{\text{TCGF}}^2]^{-1} \tilde{\mathbf{W}}_{1:7} (\tilde{\mathbf{V}}'_{1:7} \tilde{\mathbf{V}}_{1:7})^{-1} \tilde{\mathbf{V}}_{1:7} \mathbf{b}, \quad (2)$$

where T is the number of years during 1969–2018 (i.e., 50), σ_{TCGF}^2 is the variance of observed NA TCGF during this period, $\tilde{\mathbf{W}}_{1:7}$ is a matrix representing the spatial patterns of M1–M7 (Figs. 1a,c,e,g,i,k,m), $\tilde{\mathbf{V}}_{1:7}$ is a matrix representing the corresponding PCs (i.e., the gray bars in Figs. 1b,d,f,h,j,l,n), and $\mathbf{b}$ is a column vector of the regression coefficients of PC1–PC7 (i.e., the values shown in Fig. 3a). Equation (2) applies to research addressing similar issues, and details on how it is derived are provided in the appendix. Figure 4a displays the reconstructed SST pattern favorable for NA TC genesis. It is evident that SST warming in the tropical NA and cooling in the equatorial eastern Pacific tend to promote TC genesis in the NA, consistent with previous studies (e.g., Kossin and Vimont 2007; Mei et al. 2014; Patricola et al. 2014). This reconstructed anomalous SST pattern is nearly identical to the spatial map obtained from the linear

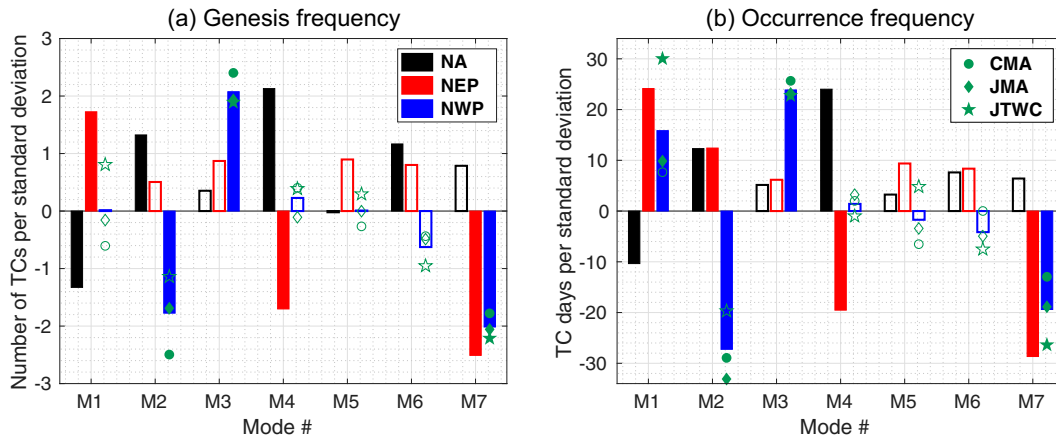


FIG. 3. The coefficients for the PCs of the seven leading SST modes from the multiple linear regression analyses of basinwide (a) TCGF and (b) TC occurrence frequency during 1969–2018 in the NA (black), NEP (red), and NWP (blue). For the NWP, dark green circles, diamonds, and stars show results from the best-track datasets produced by the CMA, JMA, and JTWC, respectively. The filled bars and symbols indicate that the regression coefficients are significant at the 0.05 level.

regression of global SSTs onto observed NA TCGF (Fig. 4b), indicating that the SST pattern favorable for NA TC genesis can be represented by several SST modes.

For the NEP TCGF, the dominant modes are ENSO and the AMO, with El Niño and a negative AMO producing more TCs (Fig. 3a). The signs of the regression coefficients show that the effects of these modes on TC genesis are opposite between the NA and NEP, in line with previous studies suggesting that TCGF in these two basins to some extent varies oppositely (Wang and Lee 2009; Collins 2010; Caron et al. 2015). The spatial pattern of SST anomalies reconstructed using the seven SST modes and multiple linear regression coefficients indicates that SST warming in the tropical eastern Pacific (particularly north of the equator) and cooling in the tropical Atlantic favor TC genesis in the NEP (Fig. 4c). As in the NA, this reconstructed SST pattern is consistent with the SST pattern obtained from the linear regression of global SSTs on observed NEP TCGF (Fig. 4d).

TCGF in the NWP is controlled by modes in association with the PMM, the interannual component of the AMO, and the global warming pattern (primarily the SST changes in the Indian and Atlantic Oceans) (Fig. 3a). Regression coefficients show that SST cooling in the tropical Indian Ocean and tropical NA and warming in the tropical central North Pacific encourage TC genesis in the NWP. This can be clearly seen in the pattern of SST anomalies reconstructed using the leading SST modes (Fig. 4e) and from the linear regression of global SSTs against NWP TCGF (Fig. 4f). These results are in line with previous studies emphasizing the importance of SSTs in all three basins of the Northern Hemisphere in influencing NWP TCGF (e.g., Zhan et al. 2011; Mei et al. 2015; Yu et al. 2016; Zhang et al. 2016; Zhan et al. 2019).

In short, tropical SST variability strongly modulates the year-to-year variations in TCGF over the NA, NEP, and NWP, but has no significant effect on those over the NIO. The SST modes that dominate TCGF differ among basins and

may have opposite effects on TCGF of the NA, NEP, and NWP. For all the three basins, a combination of the SST modes weighted by the coefficients obtained from the multiple linear regression of TCGF on the PCs of these modes well reproduces the pattern of SST anomalies that favors TC genesis. It is worth noting that TCGF in the NA and NEP is modulated by both local and remote SSTs, whereas TCGF in the NWP is primarily affected by remote SSTs.

We further repeated the above analyses on basinwide TC occurrence frequency for the NA, NEP, and NWP. The results show that tropical SST variations are more skillful in capturing the variability in occurrence frequency than the variability in genesis frequency, and account for around 61%, 67%, and 69% of the variance in occurrence frequency during 1969–2018 in these three basins, respectively (figure not shown). The modes dominating occurrence frequency are also slightly different from those controlling genesis frequency (Fig. 3b). For the NA, the AMM does not significantly affect occurrence frequency despite its importance in modulating genesis frequency. For the NEP, the global warming pattern emerges as one important factor for variations in occurrence frequency, in addition to the factors affecting genesis frequency. For the NWP, the canonical ENSO stands out as one of the dominant contributors to occurrence frequency although it has minor effect on basinwide genesis frequency.

b. Effects on the large-scale environment and local TC genesis and occurrence frequency

It has long been recognized that tropical SSTs cause the large-scale atmospheric environment to vary, and that changes in large-scale dynamic and thermodynamic environmental variables (e.g., 850-hPa relative vorticity and the magnitude of vertical wind shear) influence TC genesis (Gray 1968; Emanuel and Nolan 2004; Fu et al. 2012; Peng et al. 2012). In this section, we examine the individual effects of the seven SST modes on the large-scale atmospheric

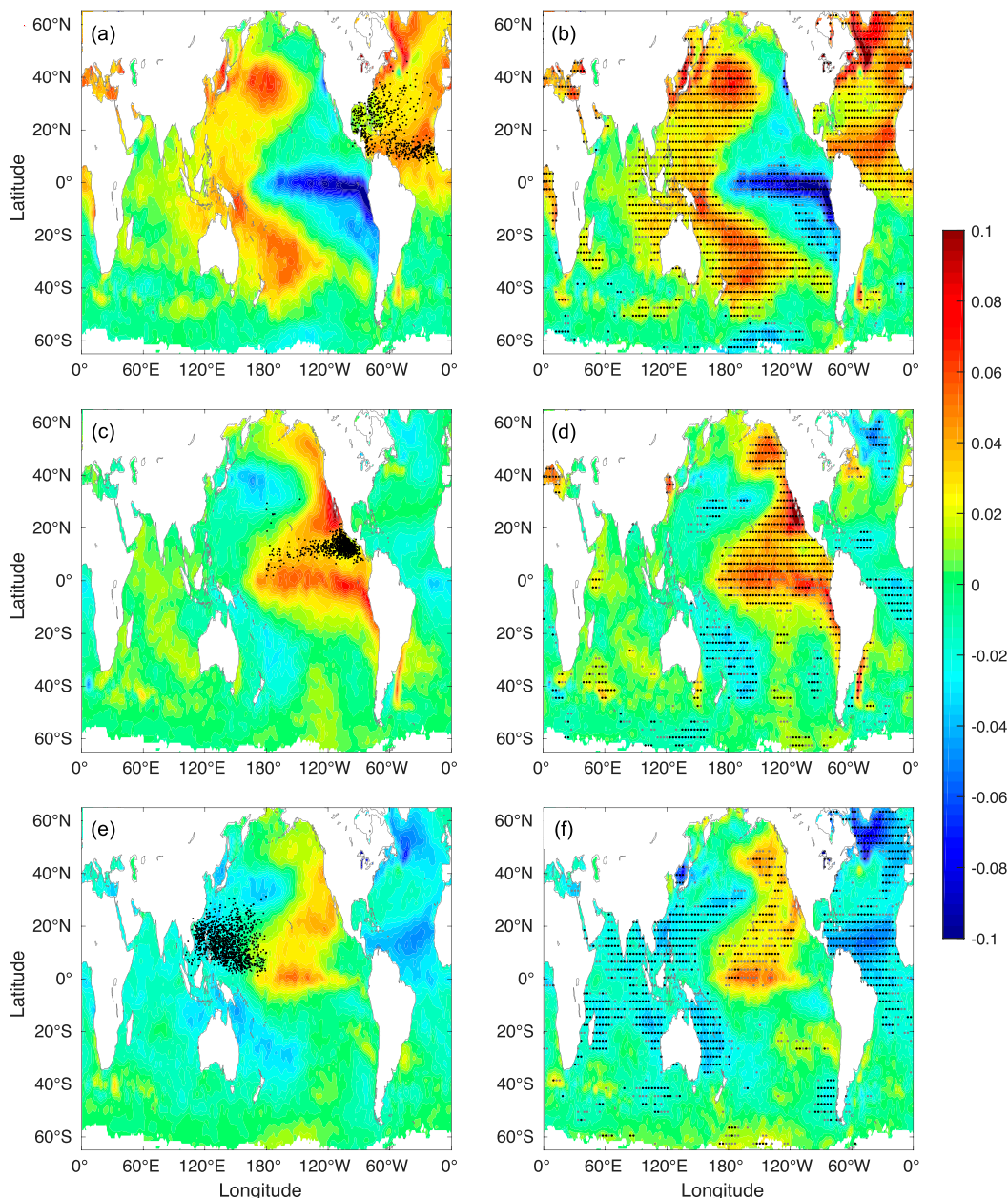


Fig. 4. The pattern of SST anomalies (shading; °C per TC) reconstructed using Eq. (2) for the (a) NA, (c) NEP, and (e) NWP. Black dots show the climatology of TC genesis location during 1969–2018. Also shown are the regressed boreal warm-season SST anomalies (°C per TC) against observed TCGF during 1969–2018 for the (b) NA, (d) NEP, and (f) NWP. Black or gray stippling denotes areas with regression coefficients significant at the 0.05 or 0.1 level, respectively.

condition and thereby on local and basinwide TC genesis and occurrence frequency in the NA, NEP, and NWP during 1969–2018 via a linear regression analysis. The large-scale environmental condition is represented by the GPI and its four components defined in section 2b based on the JRA-55 reanalysis dataset. To determine the contribution of one component (e.g., vertical wind shear) to the changes in the GPI (in response to the SST modes), we recalculated the GPI

using the original, year-to-year varying values for that component but the climatology of 1969–2018 for the other three components, following Camargo et al. (2007). This procedure was repeated for all the four components of the GPI. As a result, positive anomalies in the subplots for the four components of the GPI shown in Fig. 5 all indicate positive effects of the components on GPI, which are produced by increased potential intensity, decreased midlevel saturation

deficit, decreased vertical wind shear, and increased low-level vorticity, respectively. It is worth noting that the sum of the contributions of individual components quantified here is close but not equal to the changes in GPI, because of the nonlinearity of the GPI formula, as discussed in Camargo et al. (2007). Using the NCEP–NCAR Reanalysis 1 dataset produces very similar results on the responses of the large-scale atmospheric environment to various tropical SST modes (Fig. S2 in the online supplemental material), demonstrating the robustness of the results.

In response to a positive-phase M1, genesis potential increases in the majority of the NEP and the southeastern NWP but decreases in the western tropical NA (particularly over the Caribbean Sea) and between 20° and 40°N in the NWP (Fig. 5a). The reduction in GPI in the tropical NA is caused primarily by increased vertical wind shear and saturation deficit and tends to suppress TC genesis and development in this region (Fig. 6a). The increased genesis potential in the NEP is due mainly to reduced wind shear and increased potential intensity (Fig. 5a) and favors TC activity in this basin (Figs. 6a,b). These contribute in part to the observation that TC counts in the NA and NEP vary oppositely (Table 1, along with Fig. S3 in the online supplemental material; e.g., Wang and Lee 2009; Collins 2010). In the NWP, the increase in GPI in the southeastern portion is attributed to an increase in low-level vorticity and a decrease in wind shear, whereas the decrease in GPI in the northwestern portion is dominated by an increase in saturation deficit (Fig. 5a). This dipole pattern of GPI changes leads to out-of-phase changes in TCGF in the two subbasin regions, but a rather modest change in basinwide TC counts (Fig. 6a). Because TCs forming in the southeastern portion of the basin stay over the warm water for a relatively longer period of time and prefer recurving tracks, enhanced TC genesis in this area increases TC occurrence frequency over the entire basin except near the coastal regions around mainland China (Figs. 6a,b). These results are in line with previous studies about the effects of the canonical ENSO on NWP TC activity (e.g., Wang and Chan 2002; Li 2012). The ENSO-induced changes in genesis potential over different basins shown here are also consistent with those described in Camargo et al. (2007) despite differences in the formulation of the GPI.

A positive phase of M2 tends to increase genesis potential over the tropical NA and the subtropical and midlatitude NWP and to decrease genesis potential in the subtropical NA (Fig. 5b). In the tropical NA, reduced wind shear and enhanced potential intensity are the main contributors to the increase in GPI. The resultant increase in TCGF over the Caribbean Sea and eastern tropical NA produces a basinwide increase in TC occurrence frequency (Figs. 6c,d). In the subtropical NA, particularly off the east coast of the southeastern United States, the decrease in GPI is primarily attributed to decreased potential intensity and increased saturation deficit, and suppresses TC genesis in the region (Fig. 6c). This region, however, still experiences above-normal TC occurrence frequency because of promoted TC genesis in the tropical NA (Figs. 6c,d). In the subtropical and midlatitude NWP, the increase in genesis potential is mainly caused by increased

potential intensity and decreased saturation deficit, promoting local TC genesis. We note that the regressed map of TC genesis on PC2 shows that TC genesis is significantly suppressed in the southeastern NWP during the positive phase of M2 (Fig. 6c). This is not suggestive based on the GPI anomalies and may be attributed to the decreased low-level vorticity and increased wind shear in this region (Fig. 5b). The suppressed TC genesis in the southeastern NWP results in lower-than-normal TC occurrence frequency over the entire NWP (Fig. 6d), which is in accord with previous studies showing that the Indian or Atlantic Ocean warming may discourage NWP TC genesis (Zhan et al. 2011; Yu et al. 2016). In the NEP, the effect of strengthened wind shear counteracts that of reduced saturation deficit (Fig. 5b), leading to neutral large-scale condition and no significant changes in TC genesis (Fig. 6c). In the subtropical and midlatitude NEP, increased saturation deficit reduces genesis potential (Fig. 5b), which, however, has no effect on TC genesis because TCs rarely form in this region in the present climate (Fig. 6c).

During a positive phase of M3, genesis potential is significantly increased over the tropical North Pacific, except in the coastal regions (Fig. 5c). In the NWP, both low-level vorticity and wind shear contribute to the increase in GPI, leading to increased TC genesis and occurrence frequency over the entire basin (Figs. 6e,f). In the NEP, reduced wind shear dominates the increase in genesis potential, and meanwhile, increased saturation deficit and reduced low-level vorticity decrease genesis potential near the coastal areas of Central America. As a result, a dipole pattern of TC genesis anomaly emerges in the NEP, with more TCs forming west of and fewer TCs forming east of 115°W (Fig. 6e); the basinwide TCGF, however, does not change. The dipole of TC genesis changes also produces a dipole pattern of TC occurrence frequency in the NEP (Fig. 6f; Murakami et al. 2017a). Over the NA, M3 has no statistically significant effect on the tropical large-scale environment and TC genesis and occurrence frequency.

Mode M4, in association with the AMO, affects genesis potential primarily in the tropical NA, yielding favorable environments during its positive phase (Fig. 5d). This increase in GPI is caused by reduced wind shear and increased potential intensity, resulting in more TCs in the tropical NA and above-normal TC occurrence frequency over the entire basin (Figs. 6g,h). In the NEP, increased vertical wind shear appears to suppress TC genesis and occurrence frequency over the majority of the basin during the positive phase of M4, despite modest changes in the GPI (Fig. 5d). The weak response of the GPI to M4 in the NEP shown here and to M2 in the southeastern NWP discussed above indicate that the GPI formula in use is more sensitive to thermodynamic properties and not sensitive enough to dynamical factors of vertical wind shear and low-level vorticity in some regions, which is likely due to the complex interplay between the individual terms (Menkes et al. 2012). In the NWP, while the subtropics experiences an increase in genesis potential, changes in GPI over the tropics are small, leading to sporadic changes in local TCGF and no significant changes in basinwide TC genesis and occurrence frequency (Figs. 6g,h).

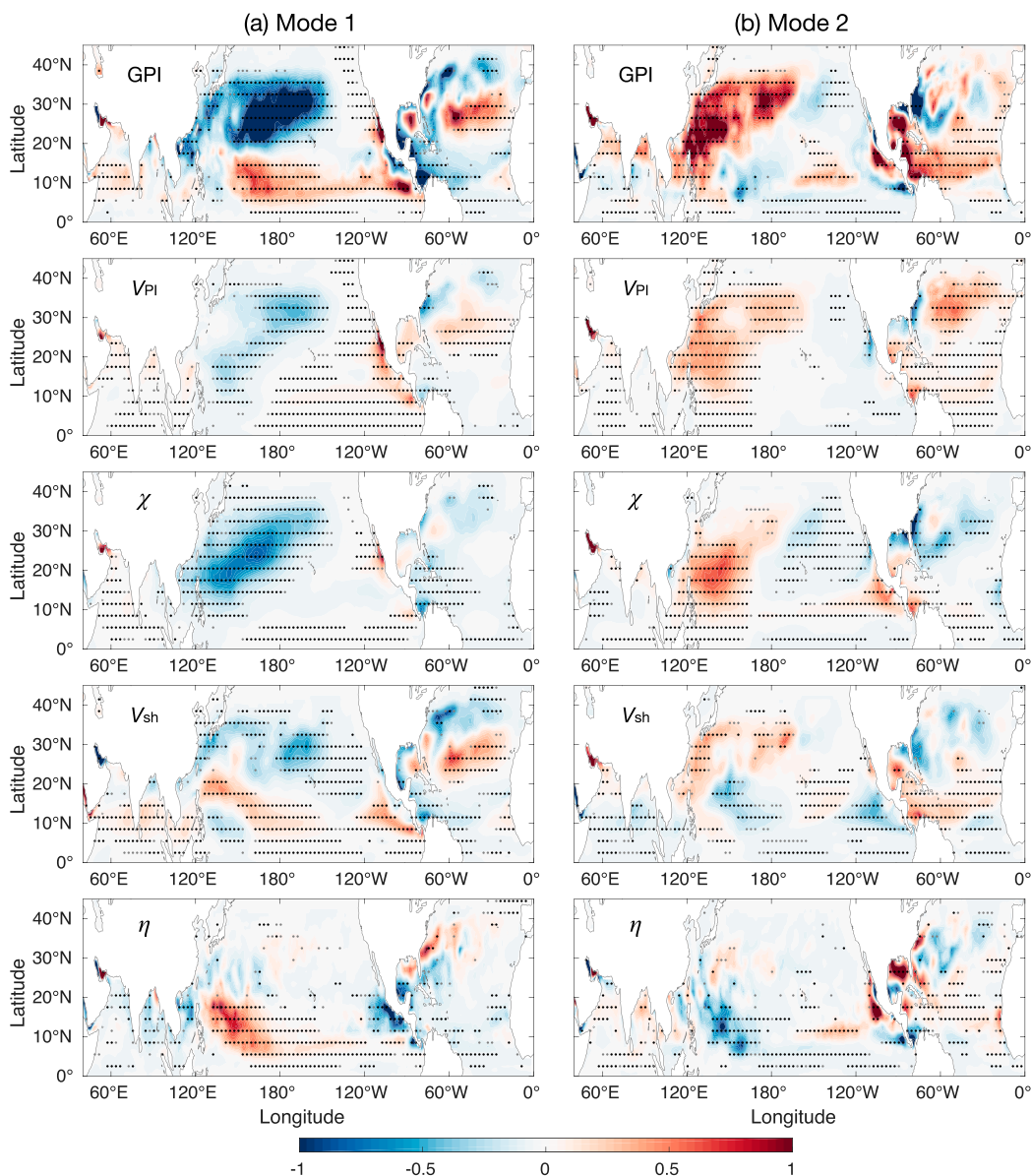


FIG. 5. The regressed anomalies of GPI and its four components (i.e., potential intensity, midlevel saturation deficit, vertical wind shear, and low-level vorticity) against the PC of (a) M1, (b) M2, (c) M3, (d) M4, (e) M5, (f) M6, and (g) M7 during 1969–2018. Black or gray stippling denotes areas with regression coefficients significant at the 0.05 or 0.1 level, respectively.

Genesis potential and TC activity over most of the Northern Hemisphere, except the NEP, are insensitive to M5 (Figs. 5e and 6i,j). In the NEP, M5 produces significant changes in the GPI through changes in vertical wind shear and low-level vorticity, and during its positive phase leads to discernible but insignificant increases in TCGF and significant increases in TC occurrence frequency over this basin (Figs. 6i,j).

During the positive phase of M6, when the tropical NA and South Atlantic respectively experience anomalous warming and cooling, genesis potential is significantly increased in the tropical NA, particularly in the eastern portion (Fig. 5f).

This increase in genesis potential can be attributed to reduced wind shear and increased potential intensity, and results in significantly increased TC genesis in this region, accompanied by an insignificant increase in TC occurrence frequency (Figs. 6k,l). The responses in other Northern Hemisphere basins are not statistically significant.

Mode M7 is associated with the interannual component of the AMO. During its positive phase, genesis potential is increased in the tropical NA but decreased in the tropical NEP (Fig. 5g). The increase in the tropical NA is contributed by both decreased wind shear and increased potential intensity,

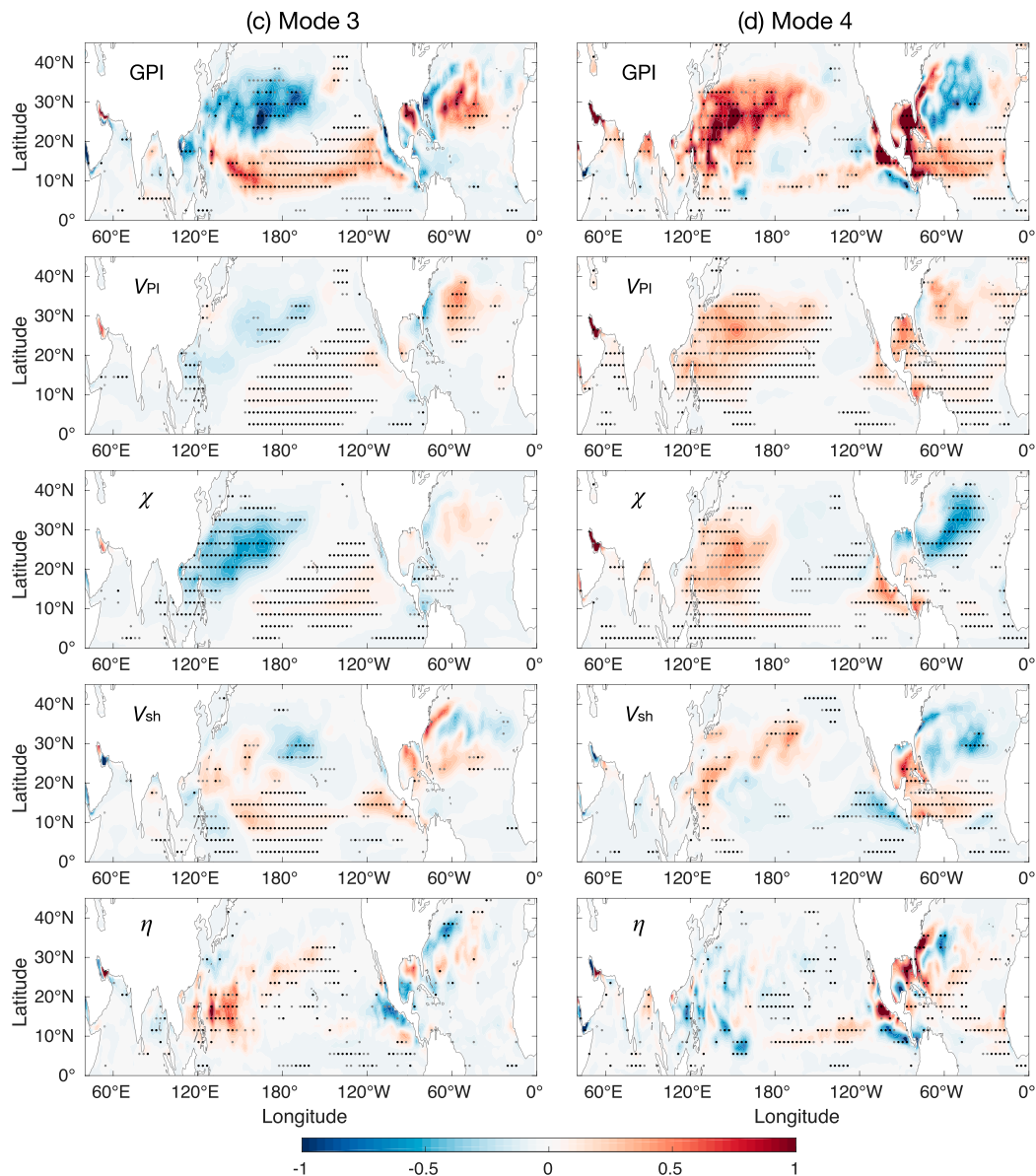


Fig. 5. (Continued)

leading to above-normal TC genesis in this basin (Fig. 6m). In the NEP, the decreased genesis potential is caused by increased saturation deficit and decreased potential intensity (Fig. 5g), which in turn can be linked to the cold anomaly north of 20°N in the NEP (Fig. 1m). As a result, the NEP experiences an inactive TC season in terms of both genesis and occurrence frequency (Figs. 6m,n). Thus, this mode also contributes to the observation of out-of-phase changes in seasonal TC counts over the NA and NEP (Fig. S3 in the online supplemental material). In addition, in response to the positive phase of M7, genesis potential tends to decrease over the NWP, particularly in the southeastern portion of the basin (Fig. 5g). The decrease in genesis potential is primarily due to increased vertical wind shear and decreased low-level vorticity (Fig. 5g), and leads to

suppressed TC genesis and occurrence in the NWP (Figs. 3 and 6m,n). The effect of M7 on NWP TC activity may be achieved through the effect of tropical NA SST anomalies on Indian Ocean SSTs and/or on the Walker circulation (e.g., Yu et al. 2016; Zhang et al. 2018). A numerical study fully exploring the effects of tropical NA SSTs with a high-resolution AGCM will shed light on this interesting topic.

c. Interdecadal modulations of tropical SST effects on basinwide TCGF

Over different periods of time, the modes that dominate the variations of TCGF over a basin may vary. In this section, we examine how the dominant modes identified in the previous sections for the period 1969–2018 change with time.

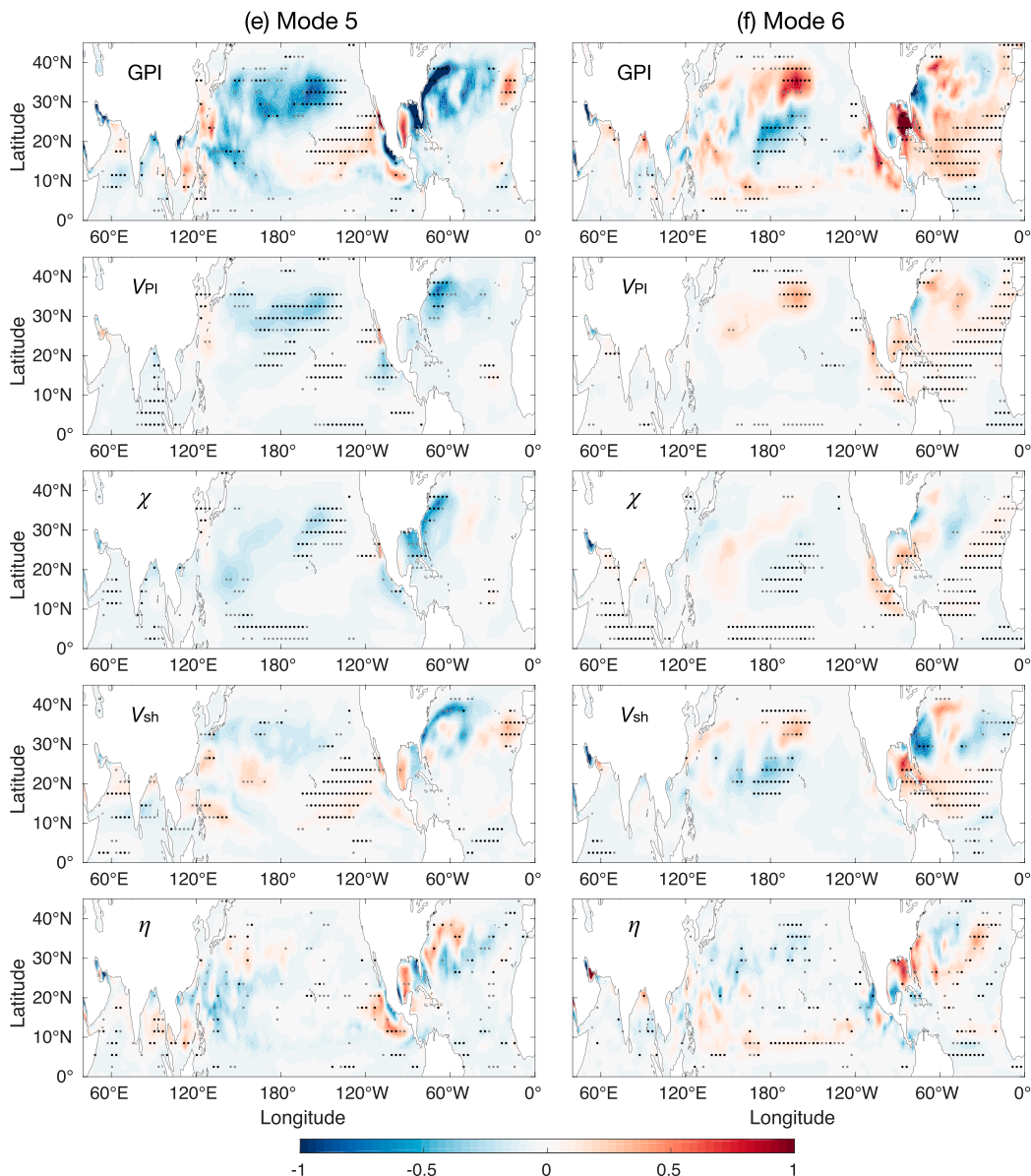


Fig. 5. (Continued)

Specifically, for each basin we apply the multiple linear regression analysis to various 30-yr periods with the starting year sliding from 1969 to 1989 (i.e., 1969–98, 1970–99, ..., 1989–2018). For each of these 30-yr periods, we obtain the regression coefficients of the leading SST modes and the correlation skill of the reconstructed TCGF.

Figure 7a shows the regression coefficients for the PCs of M1–M7 as a function of the starting year of the regression period obtained from the regression analysis of NA TCGF. Both M1 (associated with the canonical ENSO) and M4 (associated with the AMO) exert strong influences on NA TC genesis, with cooling in the central-to-eastern equatorial Pacific and warming in the tropical NA promoting TC genesis, independent of the study period. The regression coefficients of these two modes

become larger with time, and this can be attributed to both increased variance in TCGF and decreased variance in the two PCs (figure not shown). The global warming pattern (i.e., M2), which is shown to significantly affect NA TCGF during 1969–2018 in section 4a, plays a significant role only in the 30-yr periods starting before the mid-1970s. Additional multiple linear regression analyses in which PC2 is divided into interannual and decadal components suggest that such a significance is due primarily to the decadal component of this mode. The AMM (i.e., M6) modulates NA TC counts mainly after the mid-1980s.

The correlation coefficient between the reconstructed and observed TCGF varies modestly with time and is greater than 0.8 over most of the time (solid red curve in Fig. 7b), indicating the strong and stable control by tropical SSTs. We have

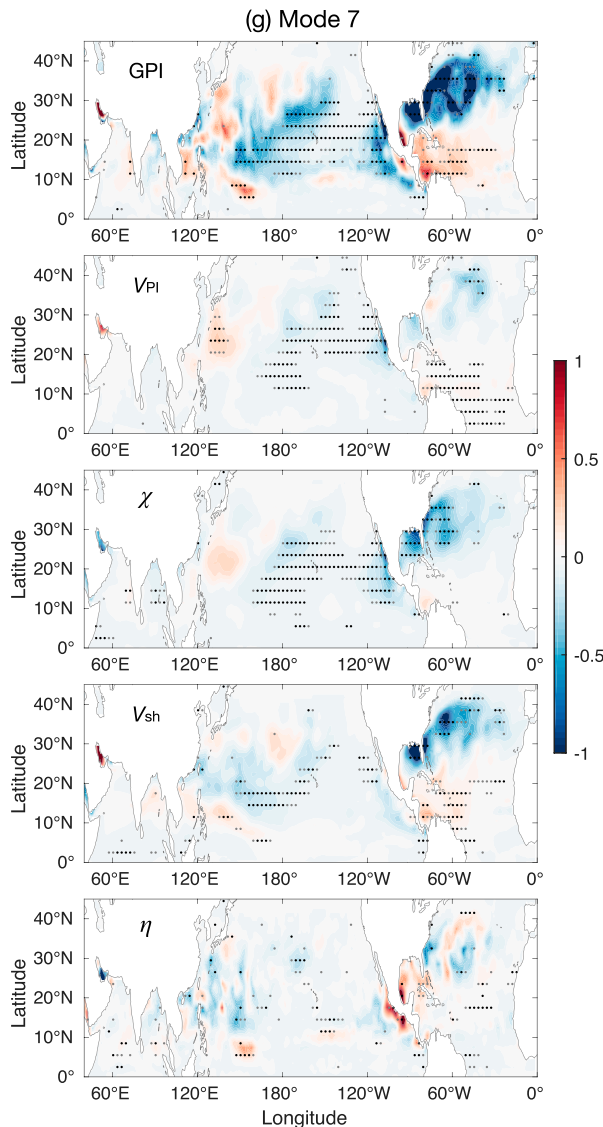


Fig. 5. (Continued)

also performed a similar sliding-window multiple linear regression analysis but using the six climate indices that correspond to the seven leading tropical SST modes: Niño-3.4, global-mean SST, the PMM, AMO, PDO, and AMM. The reconstructed TCGF (solid black curve in Fig. 7b) is also highly correlated with the observations, but the skill is lower than the reconstruction based on the leading modes of the tropical SST variations.

The modes related to the canonical ENSO and AMO (i.e., M1, M4, and M7) dominate TCGF variations in the NEP, with both El Niño and negative AMO favoring more TCs, as discussed in sections 4a and 4b. Because both modes affect NA TCGF in an opposite way in comparison with their effect on NEP TCGF, a significant correlation exists in seasonal TC counts between these two basins, particularly between the NA and western NEP (i.e., west of 112°W; Fig. S3 in the

online supplemental material; Collins 2010). The importance of ENSO in modulating NEP TCGF increases with time and has become statistically significant since the 1990s (Fig. 7c; note that the length of the sliding window is 30 years), strengthening the anticorrelation in TCGF correspondingly (Fig. S3 in the online supplemental material). As in the NA, the regression coefficients of M1 and M4 increase with time, which is also due to both increased variance in TCGF and decreased variance in PCs (figure not shown). What is different from the NA is that the mode related to the interannual AMO (i.e., M7) also modulates NEP TC counts over most of the time (cf. Figs. 7a,c).

TCGF reconstructed based on the SST modes captures 50%–75% of the variations in NEP TCGF over a 30-yr period, with the correlation skill improving with time during 1969–2018 (solid red curve in Fig. 7d). The increase in correlation skill is especially prominent in the 1990s (considering that the sliding window is 30 years long) and suggests strengthened modulations of NEP TC genesis by tropical SST variations, particularly the continuously increased importance of ENSO discussed above. The rising contribution of ENSO is further confirmed via multiple linear regression analyses with the PC of one SST mode excluded as well as a sliding-window correlation analysis of PC1 and NEP TCGF (figure not shown). The empirical model built on the six climate indices is less skillful and accounts for only 25%–50% of the variance in observed TCGF; it, however, also shows increased skill starting in the 1990s (solid black curve in Fig. 7d).

As shown in Fig. 7e, TCGF in the NWP is high during the positive phase of M3 (i.e., the PMM) throughout the entire study period. The negative phases of M7 (i.e., the interannual AMO) and of M2 (i.e., global warming) also encourage NWP TC genesis, with the former important before the late 1990s and the latter effective after the late 1990s. The TCGF reconstructed using the seven SST modes is highly correlated with the observations ($r > 0.8$), and the correlation skill is stable during 1969–2018 (solid red curve in Fig. 7f). The statistical model based on the six climate indices exhibits comparable skill but underperforms prior to the late 1980s (solid black curve in Fig. 7f).

In all three basins, excluding the modes that do not significantly affect TCGF (Fig. 3a) slightly reduces the correlation skill of the reconstructions based on multiple linear regression (cf. solid and dashed red curves in Figs. 7b,d,f). We also repeated all the above analyses but removed the trends in both TCGF and PCs of the SST modes before performing multiple regression analysis, and we obtained very similar results (figure not shown), which suggests that secular trends play a minor role. In all three basins, the modulation of basinwide TC occurrence frequency by tropical SST variations exhibits similar interdecadal changes as the modulation of basinwide TCGF and is consistently stronger than the latter (Fig. S4 in the online supplemental material).

5. Summary and conclusions

This study has investigated the influence of warm-season tropical sea surface temperature (SST) variations on Northern Hemisphere tropical cyclone (TC) genesis frequency (TCGF) using observed SSTs and TC best-track data. Specifically, we

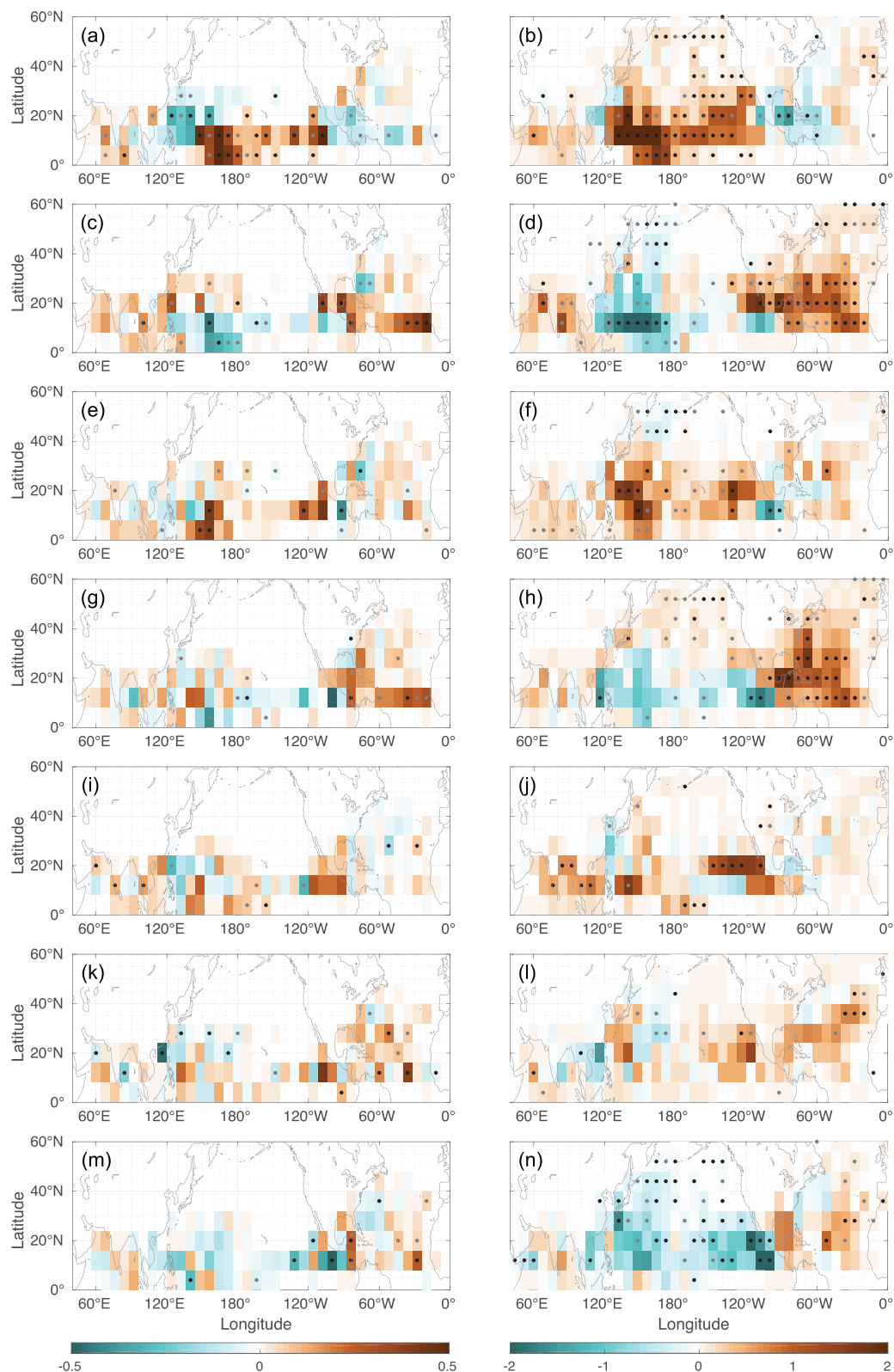


FIG. 6. The regressed anomalies of (left) TCGF and (right) TC occurrence frequency (days) against the PC of (a),(b) M1; (c),(d) M2; (e),(f) M3; (g),(h) M4; (i),(j) M5; (k),(l) M6; and (m),(n) M7 during 1969–2018. Black or gray dots denote areas with regression coefficients significant at the 0.05 or 0.1 level, respectively.

TABLE 1. Correlation coefficients r between TCGF of different basins/regions (first three rows) and variance of TCGF [$\text{Var}(\text{TCGF})$] in different basins/regions during 1969–2018. Values significant at the 0.05 level are shown in boldface type.

	NA	NEP	NWP	NA + NEP	NA + NWP	NEP + NWP	NA + NEP + NWP
NA		−0.31	−0.28			−0.35	0.21
NEP			0.44		0.11		0.73
NWP				0.17			0.70
Var(TCGF)	19.5	24.7	19.2	30.5	27.8	63.0	58.0

first extracted the leading modes of tropical SST variability during the TC peak season (June–November) via an empirical orthogonal function (EOF) analysis and then quantified the individual and combined effects of these SST modes on TCGF in each ocean basin by means of multiple linear regression analysis. We further interpreted the SST effects by analyzing the response of a genesis potential index (GPI) and its components to the identified SST modes. Potential interdecadal modulations of the tropical SST effects on TCGF are also examined. The main results are summarized as follows.

The first seven modes of tropical SST variability modulate TC activity in one or more of the basins in the Northern Hemisphere, and together account for around 80% of the variance in tropical SSTs. These modes are related to the canonical El Niño–Southern Oscillation (ENSO), global warming, the Pacific meridional mode (PMM), Atlantic multidecadal oscillation (AMO), Pacific decadal oscillation (PDO), the Atlantic meridional mode (AMM), and the interannual component of AMO, respectively (Fig. 1).

Multiple linear regression analyses show that these SST modes together capture 59%, 53%, and 56% of the variance in TCGF during 1969–2018 over the North Atlantic (NA), northeast Pacific (NEP), and northwest Pacific (NWP), respectively (Figs. 2a–c); the captured variance in basinwide TC occurrence frequency is 61%, 67%, and 69% in these three basins, respectively. These demonstrate the strong control of TC activity in the NA and North Pacific by tropical SST forcing. A linear combination of the leading SST modes weighted by the coefficients obtained from the multiple linear regression of TCGF on the PCs of these modes well reproduces the pattern of SST anomalies that favors TC genesis (Fig. 4). The SST patterns suggest that TC counts in the NA and NEP are governed by both local and remote SSTs, whereas TC counts in the NWP are affected primarily by remote SSTs. The effect of these SST modes on TC activity in the north Indian Ocean is weak (Fig. 2d), in line with previous studies based on simulations with dynamical models and analyses of GPI.

In the NA, basinwide TCGF is modulated by the AMO, canonical ENSO, global warming, and AMM: La Niña, the positive AMO and AMM, and global warming pattern favor more TCs (Fig. 3a). The first three of these four SST modes also significantly affect basinwide occurrence frequency (Fig. 3b). In the NEP, the modes dominating basinwide TCGF are in association with the canonical ENSO and AMO, with more TCs occurring during El Niño years and the negative phase of the AMO. The basinwide TC occurrence frequency is also significantly influenced by the mode associated with the PDO, in addition to those tied to ENSO and the

AMO. In the NWP, basinwide TCGF is promoted during the positive phase of the mode associated with the PMM and the negative phase of the modes associated with global warming and the interannual component of the AMO (Fig. 3a). These three modes together with the mode linked to the canonical ENSO determine basinwide occurrence frequency in the NWP (Fig. 3b).

Because TC genesis in the NWP and that in the NEP are favored under a positive PMM and a negative AMO (primarily on the interannual time scales), TC counts of these two basins have components that vary synchronously ($r = 0.44$; Table 1). On the other hand, TCGF in the NA is significantly negatively correlated with that in the North Pacific ($r = -0.35$), particularly in the NEP starting in the 1990s (Fig. S3 in the online supplemental material). The negative correlation between the NA and NEP TCGF is attributed to the fact that a La Niña state and a positive AMO favor TC genesis in the NA but discourage genesis in the NEP. As a result, the variance of the combined TCGF of these three basins (i.e., 58.0; Table 1) is smaller than the sum of TCGF variances of individual basins (i.e., 19.5, 24.7, and 19.2 for the NA, NEP, and NWP, respectively; Table 1).¹

To understand how these SST modes affect basinwide TC genesis and occurrence frequency, we further examined their effects on the large-scale atmospheric environment and on local TC genesis and occurrence frequency via linear regression analysis (Figs. 5 and 6). The large-scale environment is represented by the GPI and its four components (including potential intensity, midlevel saturation deficit, vertical wind shear, and low-level vorticity). The results suggest that a specific SST mode may affect TC genesis in different areas via different mechanisms. In general, TCGF in the NA is modulated primarily via changes in potential intensity and vertical wind shear, in line with Bruyère et al. (2012) and Mei et al. (2019). In the NEP, vertical wind shear emerges as the dominant factor for variations in TCGF, with potential intensity and

¹ To further explain this statement, let X be the NA TCGF, Y be the NEP TCGF, and Z be the NWP TCGF. The variance of $X + Y + Z$ can be written as $\text{Var}(X + Y + Z) = \text{Var}(X) + \text{Var}(Y) + \text{Var}(Z) + 2\text{Cov}(X, Y) + 2\text{Cov}(X, Z) + 2\text{Cov}(Y, Z) = \text{Var}(X) + \text{Var}(Y) + \text{Var}(Z) + 2[\text{Var}(X)\text{Var}(Y)]^{1/2}r(X, Y) + 2[\text{Var}(X)\text{Var}(Z)]^{1/2}r(X, Z) + 2[\text{Var}(Y)\text{Var}(Z)]^{1/2}r(Y, Z)$, where $r(a, b)$ represents the correlation coefficient between a and b and Cov indicates covariance. According to Table 1, $r(X, Y) = -0.31$ and $r(X, Z) = -0.28$, which lead to a negative value of $2[\text{Var}(X)\text{Var}(Y)]^{1/2}r(X, Y) + 2[\text{Var}(X)\text{Var}(Z)]^{1/2}r(X, Z) + 2[\text{Var}(Y)\text{Var}(Z)]^{1/2}r(Y, Z)$, despite $r(Y, Z) = 0.44$. This produces $\text{Var}(X + Y + Z) < \text{Var}(X) + \text{Var}(Y) + \text{Var}(Z)$.

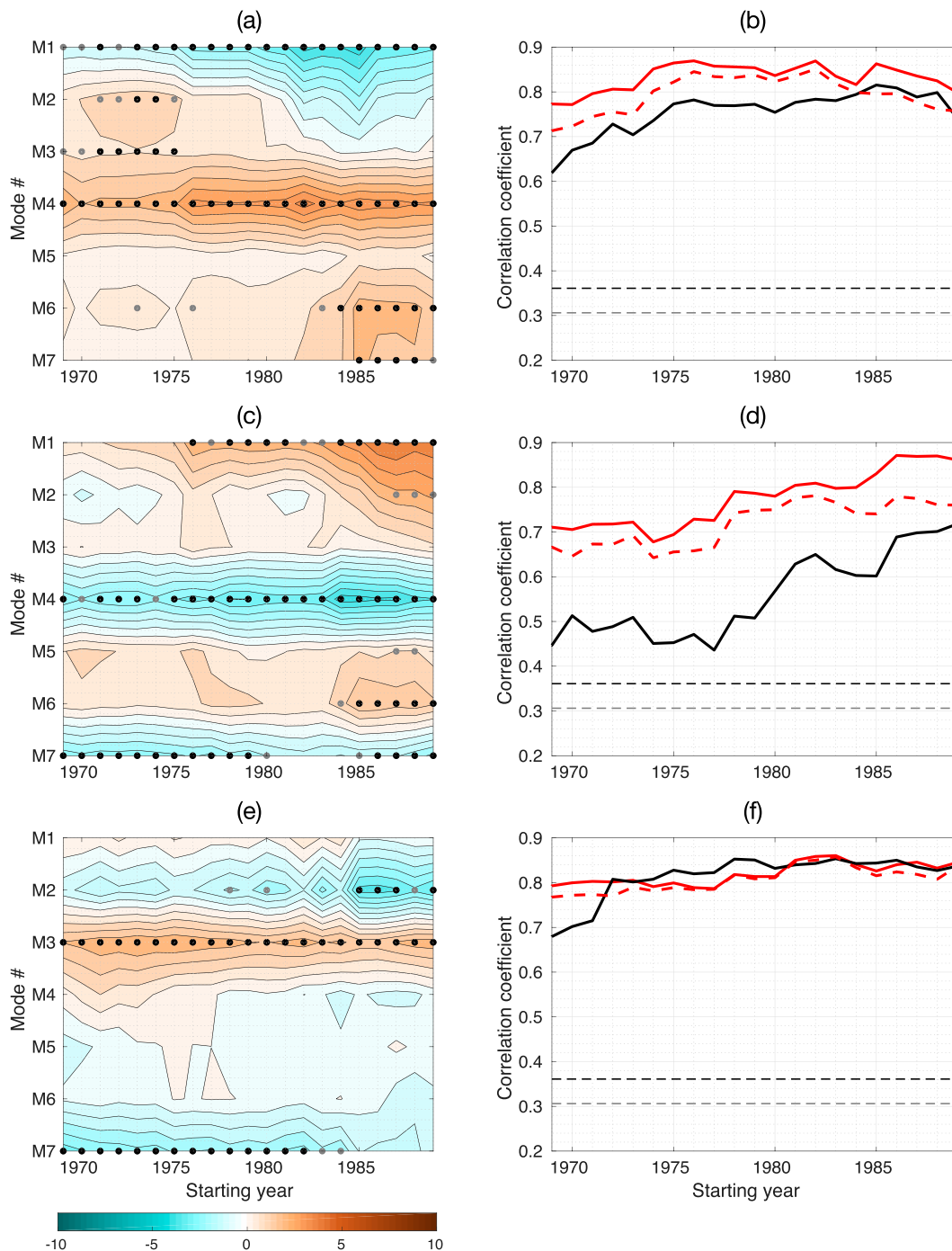


FIG. 7. The coefficients for the PCs of M1–M7 from the multiple linear regression analysis of TCGF in the (a) NA, (c) NEP, and (e) NWP during various 30-yr periods, with the starting year sliding from 1969 to 1989 (i.e., 1969–98, 1970–99, ..., 1989–2018). Black or gray dots indicate that the regression coefficients are significant at the 0.05 or 0.1 level, respectively. Also shown is the correlation skill of the TCGF reconstructed from the multiple linear regression using the PCs of M1–M7 (solid red) and using six climate indices (solid black) in the (b) NA, (d) NEP, and (f) NWP for the same 30-yr periods as in the left panels. The six climate indices in use include Niño-3.4, global-mean SST, and indices of the PMM, AMO, PDO, and AMM. Dashed red curves show the correlation skill of the TCGF reconstructed using only PCs of the modes dominant in the basin (i.e., M1, M2, M4, and M6 for the NA; M1, M4, and M7 for the NEP; and M2, M3, and M7 for the NWP). The thin dashed black or gray horizontal line in (b), (d), and (e) represents the threshold for a correlation coefficient significant at the 0.05 or 0.1 level, respectively.

midtropospheric saturation deficit making additional contributions. In the NWP, changes in low-level vorticity and vertical wind shear control TCGF in the southeastern quadrant of the basin, whereas midlevel saturation deficit is more important in the northwestern quadrant. In addition, the results echo previous studies about the opposite effects of the canonical ENSO on TC genesis in different parts of the NWP: El Niño tends to increase TC genesis in the southeastern quadrant via increased low-level vorticity and reduced vertical wind shear and suppress TC genesis in the northwestern quadrant via increased midlevel saturation deficit and decreased potential intensity.

To identify potential changes with time in the impacts of tropical SST forcing on basinwide TCGF in the NA, NEP, and NWP, we performed multiple linear regression analysis for various 30-yr periods with the starting year sliding from 1969 to 1989. It is found that the contribution of tropical SSTs to the variations in TCGF is quite stable and around 70% for a 30-yr period in both the NA and NWP (Figs. 7b,f). In contrast, tropical SSTs become more effective in modulating TC counts in the NEP, starting in the 1990s (Fig. 7d). This can be attributed to the increased contribution of ENSO (Fig. 7c) and explains the strengthened anticorrelation in TC counts between the NA and NEP (Fig. S3 in the online supplemental material).

Our work shows that warm-season tropical SSTs contribute to more than half of the year-to-year variations in TC counts over the North Atlantic and North Pacific basins during 1969–2018. We note that TC activity has been linked to tropical SST modes in preseasons (e.g., Zhan et al. 2019), a connection that needs to be investigated in future studies. Other contributing factors may include atmospheric and SST forcing from the extratropics (e.g., the North Atlantic Oscillation, Rossby wave breaking and subpolar gyre SST temperature), and forcing from the tropical land (e.g., western Sahel rainfall and dust concentration) (e.g., Landsea et al. 1999; Elsner et al. 2000; Jagger et al. 2001; Fink et al. 2010; Kossin et al. 2010; Smith et al. 2010; Dunstone et al. 2011; Caron et al. 2015; Lim et al. 2016; Patricola et al. 2018; Zhang and Wang 2019). In addition, internal variability associated with the instability of the tropical atmosphere also contributes (Chen and Lin 2011; Wu et al. 2012; Done et al. 2014; Mei et al. 2014, 2015, 2019). A thorough and accurate quantification of these factors remains challenging and may be achieved via large-ensemble simulations with high-resolution coupled models.

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APPENDIX

Derivation of Eq. (2)

Let $\mathbf{Y}_0$ be an $N \times T$ data matrix (e.g., tropical SSTs in this study; N is the number of grid points in space, T is the

length of the data in time domain, and $N > T$) and $\mathbf{Y} = \mathbf{Y}_0(\mathbf{I}_T - T^{-1}\mathbf{J}_T)$ be the centered matrix of $\mathbf{Y}_0$, where $\mathbf{I}_T$ is the $T \times T$ identity matrix and $\mathbf{J}_T$ is a $T \times T$ matrix of numeral ones. Consider the singular value decomposition (SVD) of $\mathbf{Y}$: $\mathbf{Y} = \mathbf{U}\mathbf{D}\mathbf{V}' = \mathbf{W}\mathbf{V}' = \sum_{j=1}^T d_j \mathbf{u}_j \mathbf{v}_j'$, where $\mathbf{D} = \text{Diag}(d_1, \dots, d_T)$ is an $N \times T$ rectangular diagonal matrix with the singular values in nonincreasing order, and $\mathbf{U}'\mathbf{U} = \mathbf{I}_N$ and $\mathbf{V}'\mathbf{V} = \mathbf{I}_T$. Assume that $\mathbf{1}'\mathbf{X} = 0$, where $\mathbf{1}$ is a $T \times 1$ vector of numeral ones and $\mathbf{X}$ is a $T \times 1$ vector (e.g., TCGF after removing the long-term mean in this study). Regressing $\mathbf{Y}$ on $\mathbf{X}$, $\mathbf{Y} = \beta\mathbf{X}' + \epsilon$, gives the ordinary least squares estimator of the regression coefficient β :

$$\hat{\beta} = (\mathbf{X}'\mathbf{X})^{-1}\mathbf{Y}\mathbf{X}. \quad (\text{A1})$$

Given the SVD of $\mathbf{Y}$ and considering only its first k modes, $\hat{\beta}$ is approximated by

$$\tilde{\beta} = (\mathbf{X}'\mathbf{X})^{-1}\mathbf{W}_{1:k}(\mathbf{V}_{1:k}'\mathbf{X}), \quad (\text{A2})$$

where $\mathbf{W}_{1:k}$ and $\mathbf{V}_{1:k}$ are the first k columns of $\mathbf{W}$ and $\mathbf{V}$, respectively (e.g., only the first seven modes of tropical SST variability are considered in this study).

Note that $\mathbf{V}_{1:k}'\mathbf{X}$ is the coefficient of the projection of $\mathbf{X}$ onto $\mathbf{V}_{1:k}$, and the unbiased variance of $\mathbf{X}$ is $\sigma_X^2 = (T-1)^{-1}\mathbf{X}'\mathbf{X}$. If $\mathbf{1}'\mathbf{V}_{1:k} = (0, 0, \dots, 0)_{1 \times k}$, then the variance of v_j is $(T-1)^{-1}$, $j = 1, \dots, k$. Define the normalized principal components $\tilde{\mathbf{V}}_{1:k} = (\tilde{v}_1, \dots, \tilde{v}_k) = \sqrt{T-1}\mathbf{V}_{1:k}$. Under the assumption that $\mathbf{1}'\mathbf{X} = 0$ and $\mathbf{1}'\mathbf{V}_{1:k} = (0, 0, \dots, 0)_{1 \times k}$,

$$\tilde{\beta} = \frac{1}{(T-1)^{1/2}\sigma_X^2}\tilde{\mathbf{U}}_{1:k}\left[\frac{1}{(T-1)}\tilde{\mathbf{V}}_{1:k}'\mathbf{X}\right]. \quad (\text{A3})$$

Let $\mathbf{b} = (T-1)^{-1}\tilde{\mathbf{V}}_{1:k}'\mathbf{X}$, which can be viewed as the coefficients of the projection of $\mathbf{X}$ onto $\tilde{\mathbf{V}}_{1:k}$ if $\tilde{\mathbf{V}}_{1:k}'\tilde{\mathbf{V}}_{1:k} = (T-1)\mathbf{I}_k$. In practice, a correction term is added to adjust the biases due to numerical approximations so that

$$\begin{aligned} \tilde{\beta} &= \frac{1}{\sigma_X^2}[(T-1)^{-1/2}\mathbf{W}_{1:k}]\left[\frac{1}{(T-1)}\tilde{\mathbf{V}}_{1:k}'\tilde{\mathbf{V}}_{1:k}\mathbf{X}\right]\mathbf{b} \\ &= [(T-1)\sigma_X^2]^{-1}\tilde{\mathbf{W}}_{1:k}(\tilde{\mathbf{V}}_{1:k}'\tilde{\mathbf{V}}_{1:k})\mathbf{b}, \end{aligned} \quad (\text{A4})$$

where $\tilde{\mathbf{W}}_{1:k} = (T-1)^{-1/2}\mathbf{W}_{1:k}$ (i.e., the rescaled spatial patterns of the first seven modes of tropical SST variability shown in Figs. 1a,c,e,g,i,k,m). Such an adjustment is made by noting that the ordinary least squares estimator of θ in the regression model $\mathbf{X} = \tilde{\mathbf{V}}_{1:k}\theta + \epsilon$ is $\mathbf{b} = (\tilde{\mathbf{V}}_{1:k}'\tilde{\mathbf{V}}_{1:k})^{-1}\tilde{\mathbf{V}}_{1:k}'\mathbf{X}$.

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