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**The dominant role of East Asia warming in increasing tropical
western North Pacific tropical cyclone intensity**

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The increase in intense tropical cyclone (TC) activity over the western North Pacific (WNP) has often been linked to a warming ocean¹⁻⁸. Here we show, however, that the TC intensity increasing trend in the tropical WNP during the past three decades are mainly related to the warming of the East Asian continent, especially a warming Tibetan Plateau (TP). The regional weak increasing trend of local sea surface temperature unlikely supplies the necessary energy for this increase in TC intensity. Instead, a weakened vertical wind shear (VWS) appears to be the main contributing factor. Through numerical simulations, we demonstrate that the warming TP strengthens the South Asian high-pressure system, which triggers a wave train toward the tropical WNP, subsequently modifying the upper- and lower-tropospheric zonal winds to reduce the VWS. Applying the high correlation between TC intensity and the local VWS to climate model projection results supports that TCs will likely become stronger, with a significantly increasing rate of 1.0 m s⁻¹/10 years during 2021–2050, due to a further warming of the East Asian continent. Thus, the rims of East Asia and Southeast Asia could face an increasing risk of intense typhoons.

In recent years, many super typhoons such as Haiyan (2013)⁴, Mangkhut (2018)¹¹, and Goni (2020)¹² have produced tremendous losses of human life and property damages among the costal countries of East Asia and Southeast Asia. In a warming world, the intense TC^{3,8}, the TC lifetime maximum intensity^{2,5}, and the landfalling TC intensity^{6,7} have increased in the western North Pacific (WNP). Past and projected increases in human exposure to TCs in many regions¹³ heighten the scientific and social concerns further. TC intensity is constrained by large-scale dynamic and thermodynamic factors including vertical wind shear (VWS) and moist static

energy^{1,9,14,15}. High sea surface temperature (SST) may supply more moist static energy for TC genesis and development^{1,9}, and weak VWS is unfavorable for the intrusion of environmental low-entropy air into the high-entropy reservoir of the TC inner core so that the TC warm center can be maintained^{14,15}. However, summer SST in the tropical WNP only shows a very weak or even negative trend under the global warming^{6,16-19}, which suggests that the ocean might not supply enough moist static energy for a significant increase in TC intensity. We therefore expect that changes in the atmospheric dynamic factors such as VWS could facilitate TC intensification. While many studies have linked a weakening of VWS to TC intensification^{9,11,12,14,15}, the question of the causes of such a change in VWS has not been investigated specifically. Here we show through observational diagnoses and numerical simulations that the recent warming of the East Asian continent, especially that in the Tibetan Plateau (TP), is responsible for a decrease in VWS over the tropical WNP, which subsequently lead to an increase in TC intensity.

In the WNP (100°E–180°), TCs are most active during the summer (July, August, and September; JAS) (Fig. S1). In this season, TC intensity, an average over the Joint Typhoon Warning Center (JTWC), Japan Meteorological Administration (JMA), and China Meteorological Administration (CMA) best-track datasets²⁰ (see Methods for the definition), exhibits an increasing trend in the tropical WNP (12°–24°N, 124°–143°E; defined as a key area hereafter) for the period 1988–2018 (with a relatively homogeneous time series; see Methods), with a significant linear trend of 0.2–0.6 m s⁻¹ yr⁻¹ (Fig. 1a). The annual average TC total and genesis numbers in this key area are 9.4 and 5, accounting for 56% and 30% respectively of the TC total number (16.7) over the entire WNP in JAS (Fig. S2a). More importantly, a consistent and significant increasing trend in TC intensity appears over the key area in all three best-track TC datasets, with the mean trends of 0.30 (JTWC), 0.21(JMA), and 0.27(CMA) m s⁻¹ yr⁻¹, all

significant at the 95% confidence level (Fig. 1b). This increasing trend is also characterized by a regime shift around 2003 (Fig. 1c; see Methods), with a mainly negative phase during the period 1988–2003 (the earlier period, P1) and a positive phase during the period 2004–2018 (the latter period, P2). The maximum trend occurs at the 70th percentile of TC intensity that corresponds to a wind speed of $\sim 34 \text{ m s}^{-1}$ (Fig. S2b), and the frequency of intense TCs (with the wind speed $> 33 \text{ m s}^{-1}$) significantly increases from P1 to P2 (Fig. 1d), above the 95% confidence level. The mean TC intensity (28.0 m s^{-1}) over P2 increases by about 20.2% relative to P1 (23.3 m s^{-1}), and the records of severe ($42\text{--}51 \text{ m s}^{-1}$) and super ($>51 \text{ m s}^{-1}$) typhoons in P2 (214 and 149 respectively) are 1.4 and 2.8 times of the P1 ones (152 and 53 respectively). These results indicate that intense TCs are getting stronger from P1 to P2. The TC intensification rate (IR) also shows an increasing trend. This trend increases with the quantile of IR (Fig. S2c). Because the TC number (Fig. S2a) and record (Fig. S2d) do not show significant trends, the increasing trend in TC intensity is mainly attributed to an increase in the TC IR, especially for intense TCs.

Corresponding to the increase in TC intensity, weak positive SST difference ($< 0.2^\circ\text{C}$) between P2 and P1 generally exist in the key area, while significant differences only appear to the north of the key area (Fig. S3a). This small variation in SST in the tropical WNP has been reported in previous studies^{6,16-19}. Within the key area, the additional energy from the ocean, represented by the difference in entropy deficit (see Methods) between P2 and P1 is small (200.4 J kg^{-1}) (Fig. S3b), which implies that the ocean is unlikely able to supply the necessary energy to support the increase in TC intensity. This result is distinct from hurricanes in the North Atlantic where the “storm moisture” supplied by warmer oceans fuels the increase of hurricane intensity in a warming world²¹.

The difference in VWS between P2 and P1 shows an anticlockwise anomaly in the WNP

between 10°N and 30°N, and a westerly VWS anomaly appears in the key area of the tropical WNP (Fig. 2a), which is attributed to the anomalous westerlies east of 122°E at 200 hPa and the anomalous easterlies west of 150°E at 850 hPa (Figs. 2b-c). Climatologically, the easterly VWS prevails over the tropical WNP (Fig. S4a) due to the easterly winds of 3–5 m s⁻¹ at 200 hPa west of 150°E and the weaker easterly winds of 0–3 m s⁻¹ in the lower troposphere east of 125°E (Figs. S4b and S5). Relative to the climatology, therefore, the westerly VWS anomaly indicates the weakened and strengthened easterly winds in the upper and lower troposphere, respectively, which results in a reduction of VWS, with a significant difference of -0.67 m s⁻¹ in VWS between P2 and P1 (Fig. S6a).

The ventilation index (I_v ; see Methods) can be used to evaluate the influence of the atmospheric environment on TC development and a large value of I_v inhibits TC development¹⁵. Corresponding to the weakened VWS, I_v decreases, with a significant difference of -3.6×10^{-3} between P2 and P1 (Figs. 2a and S6b). The structural changes induced by the weak VWS tend to weaken the intrusion of the environmental low-entropy air into the TC high-entropy inner core, which favors TC intensification. After removing the variation of VWS from the variability of I_v (see Methods), the rest of I_v , that is, the nondimensional entropy deficit (χ) divided by the maximum potential intensity (u_{mpi}), does not significantly decrease between P2 and P1 (Fig. S6c). It is clear that the favorable atmospheric environmental condition for TC intensification is mainly due to the contribution of the weakened VWS. The correlation coefficient between TC intensity and VWS in the key area is -0.60 during 1988–2018, significant at the 99.9% confidence level. This high correlation can be applied to an estimation of TC intensity on the basis of the environmental VWS.

We have demonstrated that a weakened VWS in the tropical WNP is the dominant factor

111 in contributing to an increase in TC intensity in the key area. So what is the mechanism that
 112 causes such a weakening of the VWS? Climatologically, upper-tropospheric easterly winds in
 113 the southeastern part of the South Asian high (SAH) prevail over the WNP (Fig. S5a). Relative
 114 to P1, a large-scale anticyclonic anomaly appears in extratropical East Asia at 200 hPa in P2,
 115 with the anomalous center in the northeastern Tibetan Plateau (TP), and the westerly and
 116 easterly wind anomalies prevail to its northern and southern sides, respectively (Fig. 2b), which
 117 indicates an eastward strengthened SAH. Large horizontal vectors (F_h ; see Methods) of the
 118 wave activity flux²² generally appear in the TP area, propagating eastward and southeastward
 119 into the tropical WNP (Fig. 2d), which indicates the propagation of wave energy into the
 120 tropical WNP and might lead to a wave train²². As a result, a cyclonic anomaly appears to the
 121 southeast of the TP anticyclonic anomaly, covering a large area from the South China Sea to the
 122 WNP, with the 200-hPa westerly wind anomalies in the key area of the tropical WNP (Fig. 2b).
 123 This anomalous pattern also appears in the lower troposphere, with an anticyclonic anomaly
 124 near 30°N, 130°E (Fig. 2c), which indicates the strengthening of the climatological subtropical
 125 high in the WNP at 850 hPa (Fig. S5b). The synchronous strengthening between the SAH and
 126 the subtropical high is associated with a strengthened zonal-vertical circulation between Asia
 127 and the North Pacific, with the upward motion around the TP and the downward motion in the
 128 central Pacific²³⁻²⁴. Meanwhile, a low-level cyclonic anomaly appears to the south of the
 129 anticyclonic anomaly center in the WNP (Fig. 2c), indicating a strengthening of the
 130 climatological monsoon trough in the tropical WNP (Fig. S5b), and the easterly wind anomalies
 131 prevail between the anomalous anticyclonic and cyclonic centers. The upper-tropospheric
 132 westerly wind anomalies and the lower-tropospheric easterly wind anomalies in the key area of
 133 the tropical WNP result in the local westerly (or weakened easterly) VWS.

The anticyclonic anomaly in extratropical East Asia between P2 and P1 is accompanied by a local warming in the middle to upper troposphere (500–250 hPa) (Fig. 2b), with a warm center of 0.8°C in the northeastern TP. Surface air temperature (SAT) also increases in the East Asian continent (Fig. 2c). The regional mean SAT significantly increases by 0.56°C over the TP area (above the topography of 1500 m) in P2 relative to P1, and it is 0.39°C for East Asia (excluding the TP area). The increasing trend of SAT in East Asia is also observed in Fig. S7. The above evidence shows that corresponding to the upward trend of the tropical WNP TC intensity, the East Asian continent is warming, with a strengthening of both the SAH and the subtropical high in the WNP. The upper-tropospheric cyclonic anomaly and the lower-tropospheric anticyclonic anomaly control the tropical (around 20°N) and subtropical (around 30°N) WNP, respectively, which together weaken the tropical WNP VWS. It is therefore hypothesized that these changes in atmospheric circulation from East Asia to the tropical WNP are forced by a warming East Asian continent, especially a warmer TP.

To demonstrate this hypothesis, we simulate an impact of the TP elevated surface heating which modifies the SAH and atmospheric circulation in the North Pacific²⁴⁻²⁶. For this purpose, the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3 (CCSM3) is conducted through changing the TP surface albedos (see Methods) and two experiments (CCSM3_Veg and CCSM3_Soil) are designed. Relative to CCSM3_Soil (a cool TP), CCSM3_Veg (a warm TP) forces a warming troposphere in extratropical East Asia (Fig. 3a) and a warming East Asian continent (Fig. 3b). Meanwhile, there are an anticyclonic anomaly over extratropical East Asia and a cyclonic anomaly over the tropical WNP in the upper troposphere (Fig. 3a), as well as an anticyclonic anomaly center around Northeast Asia and a southward cyclonic anomaly in the lower troposphere (Fig. 3b). The CCSM3 model

results are consistent with those forced by the TP surface heating with a different model²³, which demonstrates the robustness of the CCSM3 model results. Accordingly, an anticlockwise VWS circulation anomaly appears in the WNP (Fig. 3c), with the upper-tropospheric westerly wind anomalies (Fig. 3a), the lower-tropospheric easterly wind anomalies (Fig. 3b), and the westerly VWS anomalies over the key area of the tropical WNP. These simulated experiments demonstrate that the observed atmospheric environmental conditions favorable to TC development could be generated by the warming TP.

To further demonstrate that the warming TP might affect TC intensity through the above physical processes, we run two sensitivity experiments using the high-resolution Weather Research and Forecasting (WRF) for the case of typhoon Maria occurring on 3 July 2018 and passing the key area of the tropical WNP (Fig. S8a; see Methods). For the experiment with a warmer TP (WRF_veg), the model captures the atmospheric circulation anomalies similar to both the observation and the CCSM3 simulations, with a westerly VWS anomaly and a strengthened TC in the key area (Fig. S8).

Note that when the atmospheric anomalies are forced by the land surface heating, significantly positive SST anomalies to the north of the key area are also being forced, with weak SST anomalies in the key area (Fig. S9). These simulated anomalous characteristics are consistent with the observations (Fig. S3a), suggesting that the observational SST anomalies in the WNP might be forced by the warming East Asian continent. Moreover, the distribution of SST in experiment WRF_Veg is same as that in experiment WRF_Soil (see Methods), which further excludes a possible impact of the tropical WNP SST anomalies on TC intensity.

Our results demonstrate that the increasing trend of TC intensity in the tropical WNP is closely tied to a warming of the East Asian continent. This warming strengthens the South Asian

high that triggers the atmospheric circulation anomalies over the WNP, leading to a weakening of the environmental VWS and subsequently TC intensification. Compared with the impact of the warming land, the weak trend in the tropical WNP SST cannot supply the necessary moist static energy for the increase of TC intensity as in the North Atlantic²¹. Under the RCP4.5 and RCP8.5 scenarios, the tropical WNP SST does not show any remarkable increasing trend in the future 30 years (2021–2050), while the East Asian continent will continue to warm (Figs. 4a and b), consistent with the fifth phase of the Coupled Model Intercomparison Project (CMIP5) multi-model ensemble result²⁷⁻²⁹. Therefore, the warming continent will further reduce the VWS over the tropical WNP, contributing to an increase in the local TC intensity.

Based on the high correlation between TC intensity and VWS in the tropical WNP, a regression model is established (see Methods). With the three-year moving average time series of the projected VWS in the key area of the tropical WNP under the Representative Concentration Pathway (RCP) 4.5 (a medium mitigation scenario) and (RCP) 8.5 (a high mitigation scenario), the projected TC intensity will increase by $0.47 \text{ m s}^{-1} \text{ decade}^{-1}$ and $1.0 \text{ m s}^{-1} \text{ decade}^{-1}$ during 2021–2050 (significant at the 95% confidence level), respectively (Fig. 4c). This projection implies a more frequent occurrence of the tropical WNP intense typhoons with the continued East Asian warming, which will increase the risk of powerful typhoons striking the rims of East Asia and Southeast Asia such as Philippines and the southern mainland of China.

Methods

1. Observation, reanalysis, and projection datasets

Three TC best-track datasets of IBTrACS²⁰ from JTWC, JMA, and CMA are utilized to examine the consistency of TC intensity in the WNP. Following the reference³⁰, the maximum

sustained wind (m s^{-1}) is uniformly converted to a 10-minute average, in which the correction factors for the JTWC and CMA datasets are 0.88 and 0.871, respectively. Because aircraft reconnaissance of TC over the WNP stopped in 1987, which likely leads to the heterogeneity of the TC intensity time series³¹⁻³², we choose the period since 1988 in this study. The geographically-weighted regression³³ is used to produce the spatial distribution of linear trends of TC intensity in Fig. 1a. In this study, the frequency is defined as the number of TC records with a temporal resolution of 6 hr in the TC best-track datasets.

Surface air temperature at 2209 meteorological stations with continuous records from the National Meteorological Information Center of China (Fig. S7) during 1988–2018 and the monthly Hadley Centre SST data with a horizontal resolution of 1° in both latitude and longitude³⁴ are also utilized. The fifth generation of the ECMWF atmospheric reanalysis (ERA5) monthly data³⁵ with a horizontal resolution of 0.25° in both latitude and longitude is used to examine the large-scale atmospheric conditions. Because the CCSM3 model can well capture the observed atmospheric circulation changes in the East Asian-western Pacific region (including the tropical WNP VWS anomaly pattern), air temperature, wind, and SST data with Gaussian grids of 288 and 192 respectively in longitude and latitude from the CMIP5 CCSM4 simulations³⁶ are used to analyze the projection of future climate (2021–2050) under the RCP4.5 and RCP8.5 emission scenarios.

2. Statistical analysis methods

Least squares trend and composite analysis methods are applied to examine the linear trend of TC intensity and the difference between two fields, respectively. The statistical significance of composite differences and nonzero trends is assessed with the Student's t test. All the

significances in this study are at the 90% level unless otherwise stated. The nonparametric Mann-Kendall test³⁷ is employed to detect an abrupt change point of three-year running mean time series in TC intensity, showing a significant regime shift around 2003 (shown in Fig. 1c). Three-year running mean time series is utilized to establish a physical-empirical model based on the high correlation coefficient of -0.8 between TC intensity and the tropical WNP VWS by using a regression method as follows.

$$I_{TC} = -7.08 \times I_{VWS} + 93.05. \quad (1)$$

Here I_{TC} and I_{VWS} represent the predicted TC intensity and the VWS in the key area, respectively. Equation (1) is statistically checked with the F -test, significant at the 99.9% confidence level. The RMSE between the simulated and observed I_{TC} indices is 4.2 m s^{-1} during 1988–2018, accounting for 16.2% of the observed I_{TC} (25.9 m s^{-1}). Here we use the “Random-phase” test³⁸ to check the significance for the three-year running mean time series.

3. The ventilation index and entropy deficit

The ventilation index (I_v) describes the following process between TC circulation and environment. The VWS tilts TC and causes TC asymmetries. A drier and/or cooler environment relative to TC leads to a greater atmospheric entropy deficit (χ) between environment and TC, especially when the air–sea thermodynamic disequilibrium is small¹⁵. The structural asymmetries together with the entropy deficit allow low-entropy air to intrude into TC. The intrusion partially offsets the energy supplied by underlying surface fluxes, inhibiting the TC heat engine and reducing TC intensity. Thus, I_v is often used to evaluate the influence of atmospheric environments on TC development, and is defined as follows¹⁵.

$$I_v = u_{shear} \frac{\chi}{u_{mpi}}. \quad (2)$$

Here u_{shear} is the environmental VWS between 200 and 850 hPa, χ the nondimensional entropy deficit at 600 hPa, and u_{mpi} the TC maximum potential intensity (MPI). A large value of I_v inhibits TC development. Following the reference¹⁵, the ERA5 atmospheric reanalysis is applied to calculate I_v using the approach on the website (available online at <http://dx.doi.org/10.1175/BAMS-D-11-00165.2>). The entropy deficit between ocean and atmosphere, indicating the air-sea thermodynamic disequilibrium, may be used to reflect the moist static energy from the underlying ocean which supports TC development³⁹. The entropy deficit is expressed as a difference between the convective available potential energy ($CAPE^*$) of air lifted from saturation at sea level in reference to the environmental vertical profiles and the $CAPE$ of boundary layer air⁴⁰, and is defined as follows.

$$\Delta s = CAPE^* - CAPE. \quad (3)$$

The ERA5 atmospheric reanalysis is applied to calculate MPI and air-sea disequilibrium values with the code on the website (<ftp://texmex.mit.edu/pub/emanuel/TCMAX/>).

4. The energy propagation of the stationary wave

Wave activity flux is used to analyze the energy propagation of the anomalous stationary wave, aiming to identify the origin of the wave train. We use the conventional definition of the stationary wave activity flux $\mathbf{F}_s$ as follows²⁰.

$$\mathbf{F}_s = p \cos(\phi) \times \begin{pmatrix} F_x \\ F_y \\ F_z \end{pmatrix}, \quad (4)$$

where F_x , F_y , and F_z are zonal, meridional, and vertical components of $\mathbf{F}_s$, respectively.

$$F_x = v'^2 - \frac{1}{2\Omega a \sin 2\phi} \frac{\partial(v'\Phi')}{\partial \lambda}, \quad (5)$$

$$F_y = -u'v' + \frac{1}{2\Omega a \sin 2\phi} \frac{\partial(u'\Phi')}{\partial\lambda} . \quad (6)$$

All parameters in the above equations are the same as in the reference²⁰. The horizontal vector of $\mathbf{F}_s$ is defined as $\mathbf{F}_h = (F_x, F_y)$.

5. Simulations using CCSM3

The NCAR CCSM3 with a horizontal resolution of $2.8^\circ \times 2.8^\circ$ and 26 vertical levels⁴¹, which includes an atmospheric component, an oceanic component, a land model, and a sea ice model, is employed to investigate the impact of the warming TP on the atmospheric circulation in the tropical WNP during JAS. Previous studies have shown that CCSM3 can successfully simulate the impacts of Asian land heating on the Asian–Pacific climate^{42–44}. In this study, we perform two sensitivity experiments: CCSM3_Veg and CCSM3_Soil. In CCSM3_Veg, following the previous studies^{25, 44–45}, the surface vegetation type at each grid point of the TP area with a topographic height above 1500 m is set to temperate evergreen coniferous forests, which has a lower surface albedo. CCSM3_Soil is same as CCSM3_Veg but with bare soil, which has a higher surface albedo. In both experiments, CCSM3 is integrated from July 1 to September 30. Each experiment is repeated 20 times with different initial fields and the 20 ensemble mean data are analyzed. This artificial reduction of albedo on the TP results in a local surface air temperature increase (Fig. 3b), which mimics a TP warming scenario²⁵.

6. Simulations using WRF

The Weather Research and Forecast (WRF) model version 3.8.1⁴⁶, with vertical 41 levels and the model top at 50 hPa, is used to simulate the impact of the warming TP on the TC case (typhoon Maria) which occurred in the tropical WNP on 3 July 2018, developed into a super

typhoon at 21:00 UTC 5 July, and passed the key area of the tropical WNP (Fig. S8a). Details of the model settings are given in Table S1. The model domain is two-way interactive and triply nested (Fig. S8b). The outermost 45-km mesh is fixed and two nested inner meshes (with the horizontal resolutions of 15 km and 5 km) automatically move following TC. Thus the model TC is always located near the mesh centers.

To investigate the impact of the warming TP, two experiments are designed: WRF_Veg and WRF_Soil. Similar to the design in CCSM3_Veg, the landuse at each grid of the TP area (above 1500 m to the east of 85°E) in WRF_Veg is changed to broadleaf (with a lower surface albedo). WRF_Soil is same as in WRF_Veg but the landuse is changed to barren (with a higher surface albedo). The initial fields, the lateral boundary conditions, and SST in two experiments come from the ERA5 reanalysis datasets with a horizontal resolution of 0.25° in latitude and longitude at a 6-h interval³⁵. These experiments are initialized at 00:00 UTC, 2 July 2018 and are integrated for 11 days. The environmental (including the ERA5 temperature, wind, and geopotential height) nudging with a nudging coefficient of 0.0003 s⁻¹ for the outermost region is carried up to 00:00 UTC on 7 July, so that the model TC approaches the observed track. Afterwards, the environmental nudging is shut down. Following the reference⁴⁷, TC circulation is filtered from the model outputs and the filtered datasets are used to analyze environmental features. The results show that before TC enters into the key area, the TC mean kinetic energy within a 95-km radius of the TC center is equivalent between two experiments (Fig. S8c). When TC enters into the key area at 06:00 UTC 6 July, the regional mean kinetic energy remarkably increases in WRF_Veg, with a mean difference of 18.74 m² s⁻² (exceeding the 99% confidence level) in kinetic energy and a mean difference of 0.8 m s⁻¹ (exceeding the 95% confidence level) in TC intensity between the two experiments from 14:00 UTC 6 July to 14:00 UTC 9 July 2018

when both the simulated TCs are in the key area. Meanwhile, the atmospheric circulation anomalies similar to both the observation and the CCSM3 simulation are also captured (Figs. S8 d-f), including a warm East Asia in the middle-upper troposphere and the westerly VWS anomaly in the key area.

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Author contributions: XJ and ZP conceived the study and analyzed the observation and model data; CJ discussed the results and revised the manuscript; SM and XR conducted the model simulations; and YC and ZS analyzed the data and discussed the results.

Competing interests: The authors declare no competing interests.

Data availability:

The IBTrACS (version 4) of TCs during 1980–2018 are available online (at <https://www.ncdc.noaa.gov/ibtracs/>). The ERA5 monthly mean data are obtained online (<https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels-monthly-means>). The HadISST data is available at <https://climatedataguide.ucar.edu/climate-data/sst-data-hadisst-v11>. Methods and extended data in this study will be available in the online version of the paper.

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Figure 1. (a) Linear trend ($\text{m s}^{-1} \text{ yr}^{-1}$) of the mean TC intensity during 1988–2018, in which dots are significant above the 90% confidence level and the dashed box shows the key area; (b) medians (solid lines; m s^{-1}), interquartile ranges (shaded), and linear trends (dashed lines; $\text{m s}^{-1} \text{ yr}^{-1}$) for TC intensity from the JTWC (blue), JMA (green), and CMA (red) dataset in the key area during 1988–2018; (c) the TC intensity anomaly (bar, m s^{-1}) and the difference (black solid line) between the prograde and retrograde time series for 3-year running means as derived from the nonparametric Mann-Kendall test; and (d) relationship between TC frequency and intensity over the key area during the periods 1988–2003 (blue) and 2004–2018 (red). All data are for JAS only.

Figure 2. (a) Differences in the ventilation index (shaded; see Methods) and the VWS (vectors; m s^{-1} ; thick vectors are significant at the 90% confidence level) between 200 and 850 hPa between the periods 2004–2018 and 1988–2003; (b) as in (a) but for 500–250 hPa mean temperature (shaded; $^{\circ}\text{C}$) and 200 hPa wind (vectors; m s^{-1}), in which “A” and “C” denote the anomalous anticyclonic and cyclonic circulation centers, respectively; (c) as in (a) but for surface air temperature at meteorological stations (colored dots; $^{\circ}\text{C}$) and 850 hPa wind (vectors; m s^{-1}); and (d) as in (a) but for 200 hPa stationary wave F_h ($\text{m}^2 \text{s}^{-2}$). Dashed contours in (b)–(d) are for the topography above 1500 m. All data are for JAS only.

Figure 3. (a) Differences in 500-250 hPa mean temperature (shaded; °C) and 200-hPa wind (vectors; m s⁻¹) between CCSM3_Veg and CCSM3_Soil; (b) as in (a) but for SAT (shaded; °C) and 850-hPa wind (vectors; m s⁻¹); and (c) as in (a) but for VWS (m s⁻¹) between 200 and 850 hPa. All data are for JAS only.

Figure 4. (a) The projected trends of surface air temperature (shaded; $^{\circ}\text{C decade}^{-1}$) and VWS (vectors; $\text{m s}^{-1} \text{ decade}^{-1}$) between 200 and 850 hPa under the RCP 8.5 scenario during 2021–2050, in which thick vectors are significant at the 90% confidence level; (b) as in (a) but for the RCP 4.5 scenario; and (c) the observed (black) and fitted (green) TC intensity (m s^{-1}) and the projected TC intensity for the RCP 8.5 (red) and RCP 4.5 (blue) emission scenarios during 2017–2050, with their 95% interquartile ranges, in which dashed lines are for linear trends during the period 2021–2050 in the RCP8.5 and RCP4.5 scenarios and all the time series are for 3-year running means. All data are for JAS only.

475 Table S1. WRF Experiment design

Domain	Mesh 1	Mesh 2	Mesh 3
Integrated time	0000 UTC 2– 0000 UTC 13 July 2018	Same	Same
Grid point	256×178	148×148	325×325
Resolution (km)	45	15	5
Cloud microphysics	WSM6	WSM6	WSM6
PBL scheme	YSU	YSU	YSU
Land surface process	Noah	Noah	Noah
Shortwave radiation	Dudhia	Dudhia	Dudhia
Longwave radiation	RRTM	RRTM	RRTM
Cumulus convection	Kain-Fritsch	Kain-Fritsch	No

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478 Figure S1. Number of monthly TC genesis in the western North Pacific averaged among the

479 JTWC, JMA, and CMA dataset during 1988–2018.

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Figure S3. (a) Differences in SST (shaded; °C) and 850 hPa wind (vectors; m s⁻¹), between the periods 2004–2018 and 1988–2003, in which dots are significant at the 90% confidence level; and (b) entropy deficit (J kg⁻¹) averaged over the key area (solid) and its linear trend (dashed), with the difference between the periods 2004–2018 and 1988–2003 and the corresponding statistical confidence (p -value). All data are for JAS only.

499 (a)

(b)

501 Figure S4. (a) Climatological VWS (vector; m s^{-1}) between 200 and 850 hPa; and (b)
 502 longitude-height cross section of zonal wind difference (shaded; m s^{-1}) between the periods
 503 2004–2018 and 1988–2003 and the climatological zonal wind (contours; m s^{-1}) along the
 504 latitudes 12° – 24° N. All data are for JAS only.

(a)

(b)

Figure S5. Climatological JAS mean geopotential height (shaded; m) and wind (vectors; m s⁻¹) at (a) 200 hPa and (b) 850 hPa.

Figure S6. JAS anomalies of (a) VWS (m s^{-1}) between 200 and 850 hPa, (b) ventilation index (I_v), and (c) $\frac{\chi}{u_{mpi}}$ averaged over the key area and their linear trends (dashed lines), with the difference between the periods 2004–2018 and 1988–2003 and the corresponding p values.

Figure S7. Linear trend of surface air temperature ($^{\circ}\text{C}$) at meteorological stations with the total number of 2209 in China during the period 1988–2018, in which black dots are significant above the 90% confidence level.

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Figure S8. (a) Observed (black) and simulated track of typhoon Maria from 2 to 13 July 2018, in which the numbers are the mean TC kinetic energy within a 95-km radius of the TC center (m² s⁻²) at 00:00 UTC each day; (b) triply nested WRF-ARW model domains, in which the position of d02 and d03 are at the initial time; (c) time series of the simulated kinetic energy (dashed; m² s⁻²) of typhoon Maria averaged within a 95-km radius of the TC center in WRF_Veg (red) and WRF_Soil (blue), in which the short dashed lines are for times when TC enters into and leaves away from the key area; (d) differences in 500-250 hPa temperature (shaded; °C) and 200 hPa

527 wind (vectors; m s^{-1}) averaged over the period from 7 to 10 July 2018 between WRF_Veg and
528 WRF_Soil, in which “A” and “C” denote the anticyclonic and cyclonic centers, respectively; (e)
529 as in (d) but at 850 hPa; and (f) as in (d) but for VWS (vectors; m s^{-1}) between 200 and 850 hPa.
530

Figure S9. Differences in SST (shaded; °C) and 850 hPa wind (vectors; m s^{-1}) between CCSM3_Veg and CCSM3_Soil.