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Shifts of future tropical cyclone genesis date in north atlantic and north pacific basins: an ensemble modeling investigation



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Changes in the tropical cyclone (TC) seasonal cycle can have profound impacts on compound hazards associated with TCs, such as consecutive summer rainfall and TC-heatwave compound events. However, only a few studies have explored future changes in TC seasonality, and they reach discrepant conclusions. In this study, we perform a high-resolution coupled climate simulation to study the future TC seasonal cycle and investigate the mechanisms of possible changes. The model simulation shows that, under the shared socio-economic pathway 5.8.5 scenario, the mean genesis date will shift significantly to later in the season in Northeastern Pacific (ENP) and North Atlantic (NA) but shift to later or earlier depending on the subregions in Northwestern Pacific (WNP). These shifts in TC seasonal cycles are induced by seasonally asymmetric changes in TC-favorable environmental conditions, which arise from seasonally asymmetric changes in large-scale circulation patterns, including the monsoon troughs, jet stream, and tropical zonal circulation.

Tropical cyclones (TCs) are high-consequence events that cause casualties and economic losses worldwide each year. TCs are risk-laden and important not only because they are accompanied with strong winds, heavy rainfalls, storm surges, and flooding^{1–6}, but also because they can co-occur with other weather events. The compound hazards posed by multiple events simultaneously or in sequence can render higher risks than the individual events combined. For example, back-to-back TCs may result in larger damage due to the lowered coastal resilience to the later TC caused by the earlier TC^{7,8}. Consecutive rainfall related to TCs can result in devastating floods, which is best exemplified by the flood in Zhengzhou, China, in summer 2021. TCs can also be followed by heatwaves to form TC-blackout-heatwave compound events, where a TC damages a power system, causing blackouts, and a subsequent heatwave can stress the local public health system as coastal residents are deprived of air conditioning⁹.

The risk of temporal compound events involving TCs highlights a need to understand TC seasonality. Previous research shows that the chance of back-to-back TCs is influenced by TC seasonality^{7,8}. Shifting TC seasonal cycle may increase the chance of summer consecutive rainfall¹⁰, and precipitation associated with landfalling TCs can terminate droughts if they arrive during the dry season^{11,12}. In Northwestern Pacific and North Atlantic

Basins, peak TC season (late August) occurs later than the hottest season over the continents (July), making the TC-heatwave compound event rare in the historical climate^{9,13}. The change in TC seasonality may thus significantly influence the future occurrence of back-to-back TCs, TC-heatwave compound events, and hydrological extremes. Accordingly, understanding future changes in TC seasonality is critically important for projecting future risks and social impacts associated with TCs.

Although there have been ample studies related to TC climatology, few of them have focused on the future shift in TC seasonality. Previous studies reported earlier onset of the NA hurricane season based on historical record¹⁴ and a lengthened TC season in the future¹⁵. A recent study¹⁰ also reported an early-shifting trend of the occurrence time of intense TCs in the historical record and, by examining oceanic conditions simulated by climate models, deduced that there will also be a seasonal advance of intense TCs in the future. However, another recent study¹⁶ reported a projected delay in the onset of the future TC season, based on atmospheric-only climate modeling with prescribed sea surface temperatures, and it attributed such changes to changes in the Intertropical Convergence Zone (ITCZ). The discrepancies between previous studies invite further research using advanced modeling frameworks to examine

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the mechanisms that control changes in TC seasonality in the future climate.

Here we provide a more reliable projection of future TC seasonality in the Northwestern Pacific (WNP), Northeastern Pacific (ENP), and North Atlantic (NA) basins using high-resolution fully coupled large-ensemble climate model simulations. We focus on these three specific basins as TCs, their impact densely populated and economically developed regions, and share similar seasonality in historical observations. We employ the Geophysical Fluid Dynamics Laboratory (GFDL)-developed Seamless System for Prediction and Earth System Research (SPEAR, with high fidelity of TC climatology simulation¹⁷, see details in “Method”) model to simulate TCs under historical and future Shared Socioeconomic Pathway 5 8.5 (SSP5 8.5) conditions. Two numerical experiments over the period of 1979–2100 are conducted. We include both anthropogenic (greenhouse gases, anthropogenic aerosols, etc.) and natural forcings in one experiment (All_forcing), and only natural forcings in the other experiment (Nat_forcing). For each experiment, thirty ensemble members are included to provide a robust estimation of future changes in TC genesis date. We find that in ENP and NA, the mean genesis date of TCs is projected to shift 1–2 weeks later by the end of this century, while mean genesis date of TCs in WNP presents different trends depending on subregions. We attribute the changes to the shifts in the Indian-East Asian monsoon trough, equatorward extension of the midlatitude jet, and the changes in the tropical zonal circulation. The large ensemble of the SPEAR simulations can be used as a main TC hazard assessment dataset, and a thorough investigation of TC seasonality shift in this dataset will enhance interpretability of the assessment of TC-involved compound hazards in future research.

Results

Change in mean TC genesis dates

In this study, we focus on the mean genesis date of all the TCs generated in a basin as a characteristic of TC seasonality. We choose to analyze the mean genesis date (rather than the onset of the TC season) as it is more reasonable to explain the change of this variable based on the change of averaged environmental parameters, which simplifies the analysis. Ensemble results

from SPEAR simulations indicate that the mean TC genesis date will shift to late season by the end of the 21st century in ENP and NA in the All_forcing experiment (Fig. 1). In 2100 in ENP and NA basins, the ensemble mean genesis date in the All_forcing experiment is 252 and 251 (September), respectively, while in the Nat_forcing experiment the mean date is 236 and 238 (August), respectively. The mean genesis dates in the control experiment in ENP and NA are 235 and 237, respectively. To avoid influence from extreme early or late TCs, medians of the genesis date are also calculated. In 2100 in ENP and NA basins, the average of the medians of genesis date among the 30 ensemble members is 257 and 254, which are also around 2 weeks later compared to the medians in 1979 (240 for ENP and 239 for NA). Thus, the anthropogenic forcing causes around 2 weeks shift of mean TC genesis date into late season in ENP and NA. Such late-season shift is not simulated by the Nat_forcing experiment for the selected basins. However, in the WNP basin, the mean genesis dates in the Nat_forcing and All_forcing experiments have no significant difference or temporal changes, nor do the medians.

To understand the late seasonal shift of the TC genesis date, we investigate the changes in storm frequency and environmental conditions in early and late seasons. In this study we define May to July (MJJ) as early TC season and September to November (SON) as late TC season, as the historical mean genesis date (averaged from 1979–2014 in All_forcing experiment) in the three basins are all in late August (Fig. S1), and by defining this way we avoid the possible changes in the peak season. The genesis locations in each basin have no significant differences between 1979–2014 and 2065–2100, except for NA during early season (Fig. S2). However, the genesis frequency in the late season decreases (from 5.34 to 4.35 storms per year) less than in the early season (from 2.11 to 0.74 storms per year) in NA (Fig. S2), and for ENP the genesis frequency in the late season increases (from 6.97 to 8.64 storms per year) while the genesis frequency in the early season decreases (from 3.85 to 2.95 storms per year). The temporal asymmetry of the change in storm frequency explains the shift in the mean genesis date in the ENP and NA basins.

Detailed examination of the mean genesis date change in WNP shows that there are three sub-regions that show different changes in TC mean

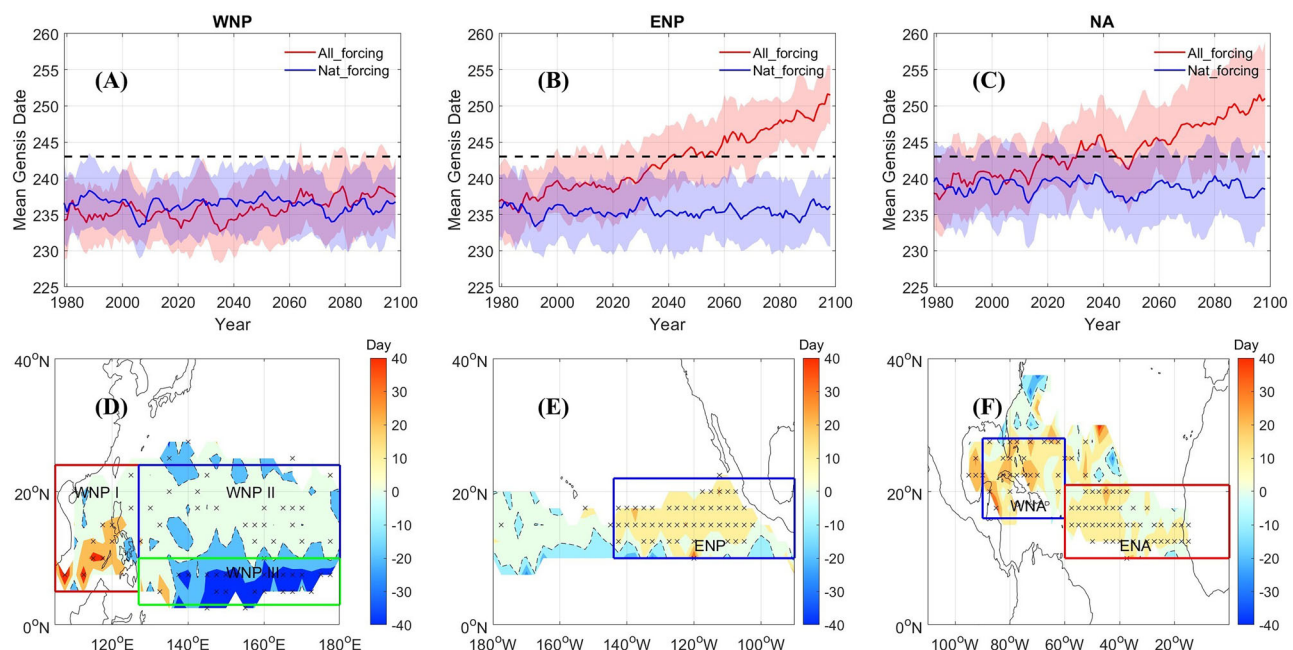


Fig. 1 | Mean TC Genesis Date. Basin-average mean TC genesis dates from 1979 to 2100 in (a). WNP, (b). ENP, and (c). NA; and the change in mean genesis date in All-forcing experiment from 1979–2014 to 2065–2100 in (d). WNP; (e). ENP; (f). NA. For (a, b, c), the solid blue and red curves show the ensemble mean in the Nat_forcing and All_forcing experiments, respectively. The shading indicates the spread

of the 30 ensemble members. The dashed black line separates August and September. For (d, e, f), the cross indicates the locations where the change is statistically significant using two-sample *t*-test under 5% significance level. The dashed black contour indicates the level of no change. The rectangles show the sub-regions used in the following sections for environmental condition analysis.

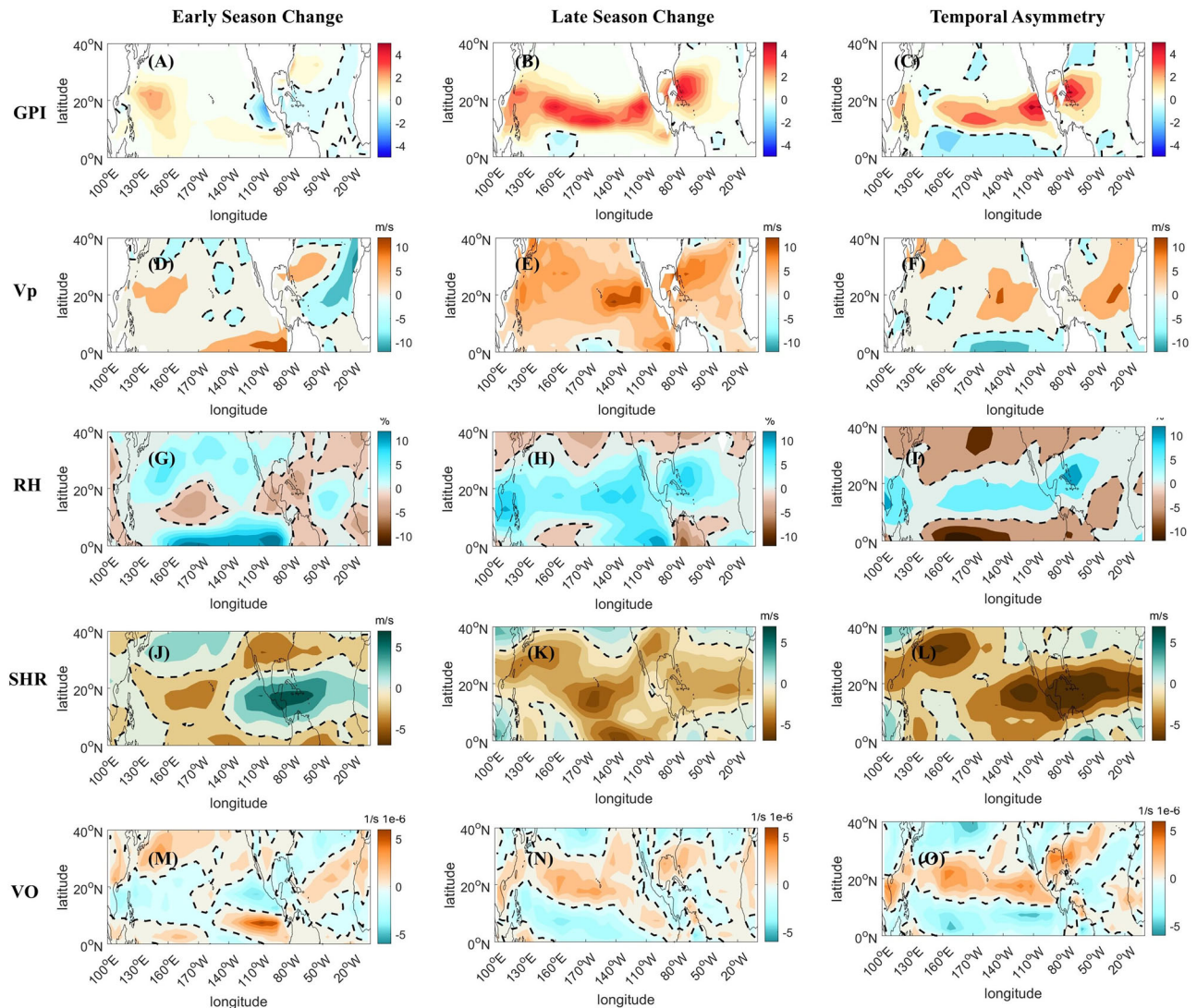


Fig. 2 | Projected changes (between 2056–2100 and 1979–2014) in the ensemble mean season-average TC-genesis-related environmental parameters. A–C GPI, D–F Vp, G–I RH, J–L SHR, and M–O VO in early season (May–July, left column)

and late season (September–November, middle column), and the temporal asymmetry of the changes (changes in late season minus changes in early season, right column). The dashed black contour indicates the shading level of zero.

genesis date (Fig. 1d). In South China Sea (WNP I), TC mean genesis date will shift to later season by more than a month in 2065–2100 compared to 1979–2014; In open WNP ocean north of 10°N (WNP II), there is a weak but statistically significant late season shift within 10 days; In open WNP ocean south of 10°N (WNP III), TC genesis will shift to earlier season by around a month. The detailed time series of mean genesis date change in the three sub-regions are shown in Fig. S3. The overall unchanged TC mean genesis date in WNP can be attributed to the cancel-out effect by the changes in the three sub-regions. In ENP, most of the late season shift is observed east of 140°W (hereafter ENP). In NA, there are mainly two sub-regions where statistically significant late-season shift of TC mean genesis date is observed, namely the Gulf of Mexico and East of the Caribbean (hereafter WNA), and the East of NA basin (hereafter ENA). We separate the NA basin into two sub-regions as we will discuss the different physical mechanisms for the late-season shift of TCs in these two regions. The following sections will focus on understanding the changes in the large-scale features in these sub-regions and examining how they favor the simulated shifts of TC mean genesis date.

Temporal asymmetry of genesis potential index (GPI) changes

The GPI is calculated using Eq. (1) following ref. 18 to understand how the changes in large-scale environmental conditions contribute to the simulated

changes in TC seasonality in the All forcing experiment. We chose to use the Emanuel–Nolan GPI in Eq. (1) (rather than the dynamic GPI) to make better connection with both previous studies in TC seasonal cycles^{19,20} using the energy GPI and the synthetic storm model downscaling²¹ which showed high correlation with the energy GPI²² and will be applied later in this study to confirm the findings reached by the SPEAR model.

$$GPI = (1.0 + 0.1 \times SHR)^{-2.0} \left(\frac{RH}{50} \right)^3 \left(\frac{Vp}{70} \right)^3 \times |VO \times 10^5|^{1.5} \quad (1)$$

where Vp is the potential intensity, RH is the 600 hPa relative humidity, SHR is the deep level vertical wind shear (200 hPa wind – 850 hPa wind), and VO is the vertical component of 850 hPa absolute vorticity. GPI reasonably captures the seasonal cycle of TCs in the selected three basins in the SPEAR simulations (Fig. S1). The temporal asymmetry of changes (the change in late season minus the change in early season) in GPI (Fig. 2a–c) is large, especially from 160°E to 40°W.

To understand why there are temporal asymmetries of changes in GPI in the selected sub-regions, we examine the changes in each parameter (Fig. 2d–o). We also performed a sensitivity analysis by changing one variable at a time to examine which environmental factor(s) have strong impact on the

Table 1 | Contribution of each environmental parameter in the seasonal asymmetry (SON–JJA) of changes in log(GPI)

	WNP I	WNP II	WNP III	ENP	WNA	ENA
log(GPI)	+0.45	+0.31	-0.25	+1.05	+1.00	+0.48
V_p	+0.07 (+15.6%)	+0.02 (+6.4%)	+0.01 (-4.0%)	+0.21 (+20.0%)	+0.11 (+11.0%)	+0.26 (+54.2%)
RH	+0.30 (+66.7%)	+0.23 (74.2%)	-0.00 (-0.00%)	+0.36 (+34.3%)	+0.54 (+54.0%)	-0.04 (-8.3%)
SHR	+0.07 (+15.6%)	+0.05 (+16.1%)	-0.01 (+4.0%)	+0.45 (+42.9%)	+0.33 (+33.0%)	+0.24 (+50.0%)
VO	+0.01 (+2.2%)	+0.01 (+3.2%)	-0.25 (+100.0%)	+0.03 (+2.9%)	+0.02 (+2.0%)	+0.02 (+4.2%)

The number shows the change of log(GPI) (first row) caused by the change of each individual variable in each sub-region. For each sub-region (column), the parameter that has the highest impact are bolded, and the parameters that have greater than 1/3 of the highest impact are shown in blue.

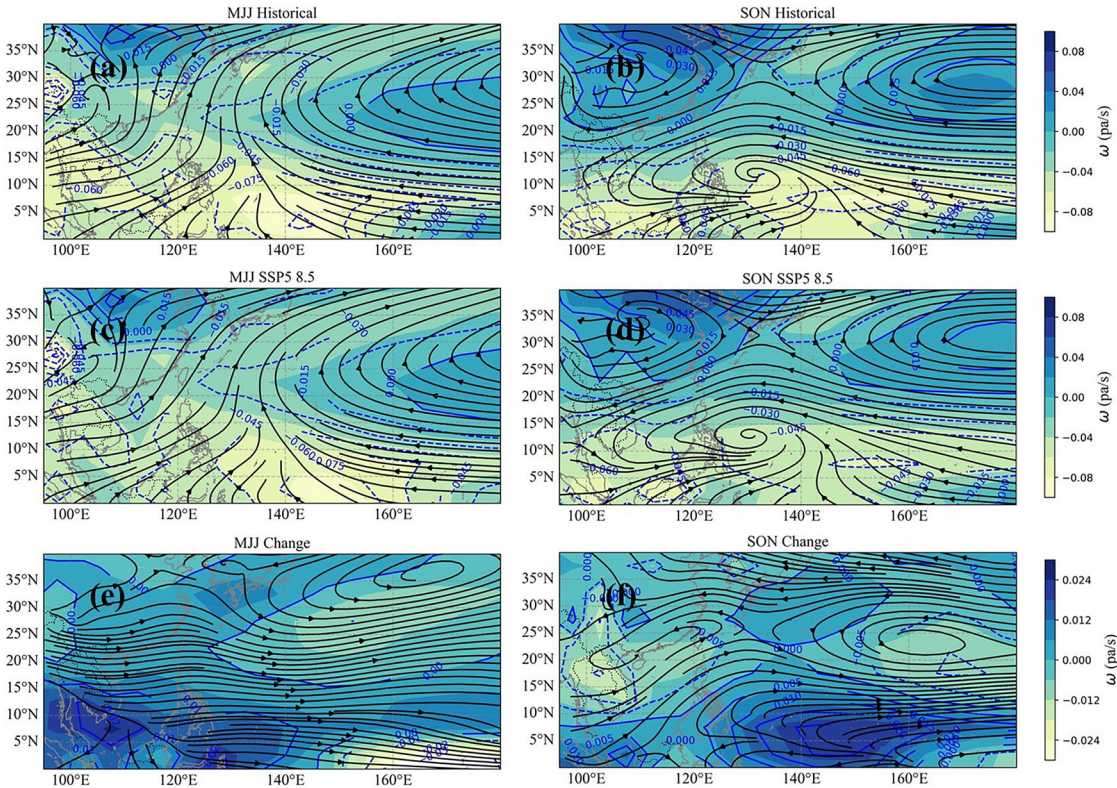
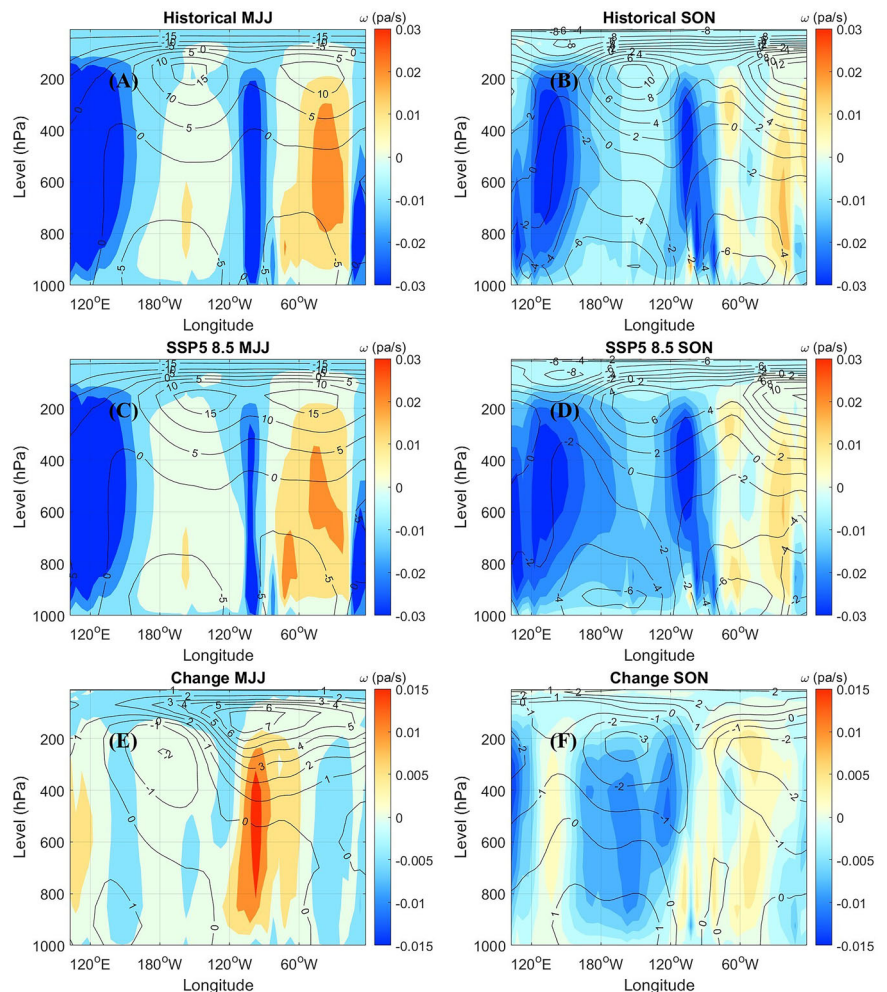


Fig. 3 | Monsoon trough and the associated deep convection. The solid arrow curves are the 850 hPa streamline, and the contours (and the color) display the 500 hPa p-coordinate vertical velocity. **A, C, E** show the circulation during early season (MJJ) in the historical condition and SSP5 8.5 condition, and the change,

respectively. **B, D, F** show the circulation during late season (SON) in historical conditions, SSP5 8.5 conditions, and the change, respectively. The solid green curves in (**A, C**) show the locations of the monsoon troughs, and the dashed green lines in (**B, D**) show the locations of the cyclonic circulation centers.

Fig. 4 | Zonal circulation averaged between 10°N–25°N. Zonal circulation averaged between 10°N–25°N in early season (MJJ) (a)(c)(e) and late season (SON) (b)(d)(f). The contour shows the zonal wind speed, and the shading shows the p-coordinate vertical velocity. **A, B** Averaged zonal circulation between 1979–2014 during MJJ and SON, respectively. **C, D** Averaged zonal circulation between 2065–2100 during MJJ and SON, respectively. **E, F** Change of zonal circulation from 1979–2014 to 2065–2100 for MJJ and SON, respectively.



temporal asymmetry of the changes in the logarithm of GPI (Table 1 and Method). The contribution calculated in Table 1 is the average influence across the whole respective sub-regions. Changes in the mid-level relative humidity explain most of the late season shift of TC genesis date in WNP I, WNP II, and WNA, and have relatively significant influence in ENP. Changes in deep-level wind shear are most important to the late-season shift in ENP and contribute to the late-season shift of TCs in WNA and ENA. Changes in potential intensity are important for the late season shift in ENP and ENA, while the change in low-level vorticity is the sole dominant factor for the early season shift of TC geneses in WNP III, as the impacts of V_p , RH , and SHR are low and cancel each other. In the following sections, we will examine the large-scale circulation features that influence the changes of these parameters in their high-impact sub-regions.

Change in the monsoon trough and its relationship with TC seasonality change

In WNP, the changes in TC seasonality are mostly related to mid-level relative humidity change (WNP I, WNP II) and low-level vorticity change (WNP III). Both factors appear to be related to the changes in monsoon trough (MT). Deep convection associated with MT increases the vertical transport of moisture and thereby modulates the mid-level relative humidity, and the location of MT influences the low-level vorticity. Previous research²³ studied the shifts of MT in the future and showed that future MT will extend eastward and intensify, but they didn't investigate the changes separately for different TC seasons. We found that, in the early season, MT (Fig. 3) is around 10°N during 1979–2014 and will extend further eastward in the future. The eastward extension will weaken the deep convection in

WNP I but enhance convection and relative vorticity in WNP III. In the late season, there will be a cyclonic circulation change located northwest of the historical circulation center (clearer in Fig. 3f), which enhances the convection in WNP I and WNP II. The circulation changes associated with MT will result in the increase (decrease) in low-level vorticity in WNP III in the early (late) season, which explains the early season shift of TC geneses in this sub-region. The increased deep convection associated with MT in late season will result in significant increase in mid-level relative humidity in WNP I and WNP II, which results in the late season shift of TC genesis date in these sub-regions.

Change in zonal circulation in 10°N–25°N and its relationship with TC seasonality change

Changes in TC seasonality in ENP and WNA are partly related to relative humidity changes, which are influenced by changes of zonal circulation in 10°N–25°N (Fig. 4) in these sub-regions. Several previous studies have used historical observation and climate models and shown that the zonal circulation is weakening^{24,25}. Our simulations show that, during the early season, the zonal circulation will be weaker in the future in ENP and WNA (centered around 100°W, Fig. 4). The weakening of the zonal circulation will result in lower relative humidity (resulting from weakening of deep convection, Fig. S5), which contributes to fewer TC geneses in these sub-regions (Fig. S2). However, in SON, the zonal circulation is significantly enhanced in ENP, which leads to more TCs genesis in late season in ENP (Fig. S2). Convections associated with the zonal circulation in WNA are slightly increased or roughly unchanged in the late season. Considering the weakening of the convection associated with the zonal circulation in WNA

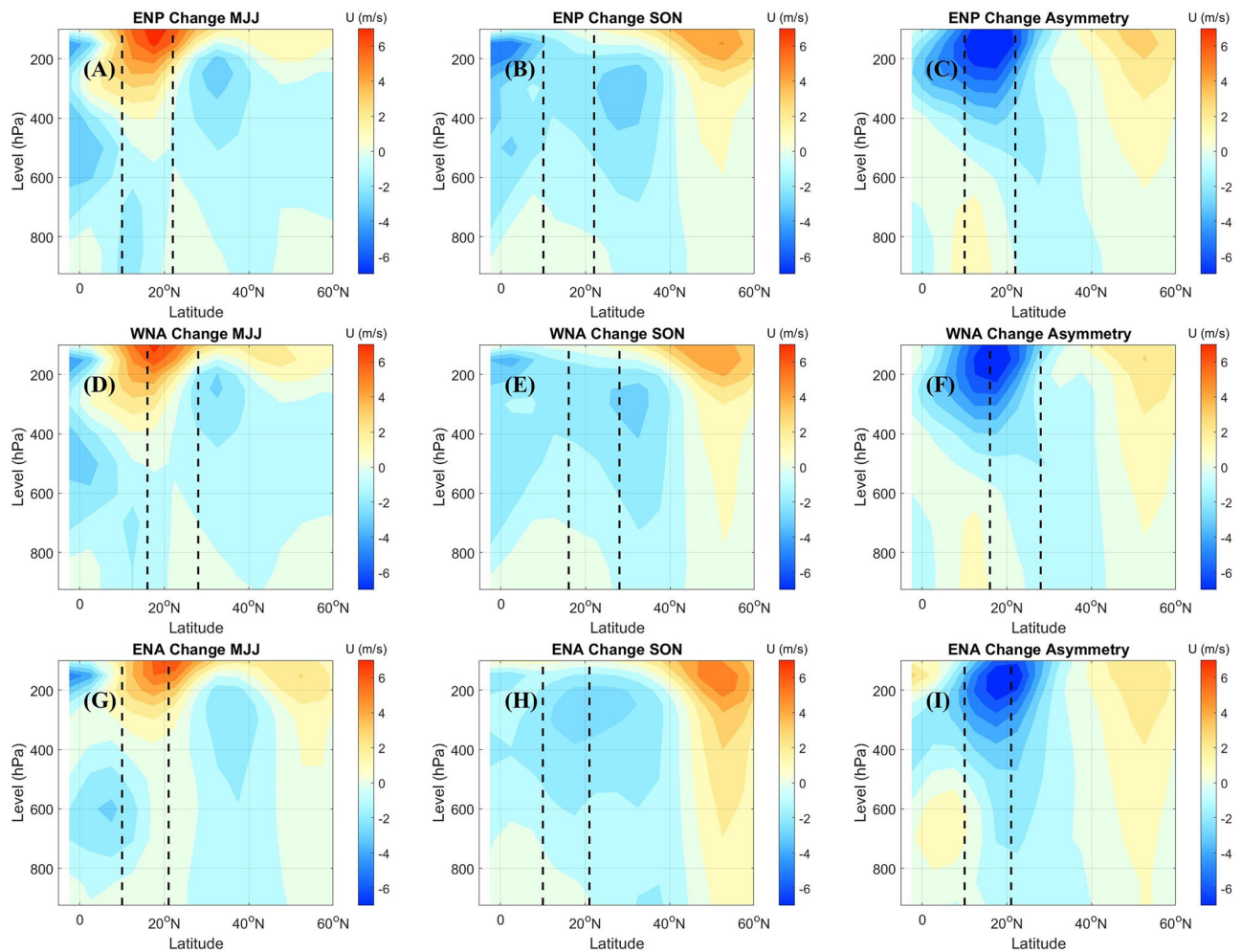


Fig. 5 | Zonal wind speed in ENP and ENA. Change in zonal wind speed in ENP (A, B, C), WNA (D, E, F), and ENA (G, H, I). The left column is the change in early season (MJJ), the middle column is the change in late season (SON), and the right

column is the asymmetry of the change (change in SON minus change in MJJ). The dashed black lines show the boundaries of sub-regions defined in Fig. 1.

during MJJ, the seasonal asymmetrical changes in the zonal circulation will result in higher increase in relative humidity in the late season in the future, leading to the late season shift of TC genesis in WNA. The change in the intensity of the convection is likely related to the shifts of ITCZ (represented as high value of monthly averaged precipitation) location (Fig. S6). For region between 120°W and 85°W, the ITCZ extends less north during MJJ in the future but extends further north during SON, which explains the weakening of convection in this region.

Change in jet stream and its relationship with TC seasonality change

Change in deep-level wind shear is important for TC seasonality change in ENP, WNA, and ENA, and the change in shear is controlled by the change in zonal wind shear (Fig. S4). Such changes in wind shear are related to the shift of Jet Stream. Compared to the Jet Stream in the historical condition (1979–2014, Fig. S7), in ENP, WNA, and ENA, the core of high-level jet will shift equatorward during the early season but shift poleward during the late season (Fig. 5). In the early season, the core of the Jet Stream will enhance in the TC genesis regions, resulting in stronger wind shear which prohibit TC genesis. In the late season, the core of the Jet Stream will shift away from the regions of TCs formation, which favors TC genesis by decreasing the wind shear. Our finding is consistent with previous studies indicating that overall anthropogenic forcings will lead to poleward shift of the midlatitude jet in the future²⁶ in boreal fall and winter due to greenhouse-gas-induced Arctic ozone depletion²⁷.

Change in sea surface temperature and its relationship with TC seasonality change

Change in potential intensity is the leading factor (secondary factor) for late season shift of TC genesis in ENA (ENP). Previous idealized radiative-convective-equilibrium simulations²⁸ found that V_p will increase as sea surface temperature (SST) increases. In the SPEAR simulations, SST change (Fig. 6) has a similar temporal asymmetry as in V_p ; SST will increase more in the late season than in the early season by around 1.5 K in the tropical region from 170°E to 40°W. Such temporal asymmetry in SST changes does not extend to the tropical NWP, although it extends to subtropical NWP.

A detailed examination of why SST changes in ENA and ENP present seasonal asymmetry is presented in the Supplementary, and here we briefly present a summary. Energy budget analysis shows that the higher SST increase in ENA and ENP during the late season (SON) than during the early season (MJJ) is a result of higher increase in energy storage (Table S1, Table S2) in June to August (JJA) than in February to April (FMA). Less increase in surface evaporation (and latent heat flux from ocean to atmosphere) in JJA than FMA is the key reason for more increase in energy storage in peak summer (Table S1, Table S2). In ENP, the less increase in surface evaporation during JJA than during FMA is a result of the stabilization of the atmospheric boundary layer during JJA, which reduces the efficiency of evaporation over sea surface (Table S3). In ENA, the less increase in surface evaporation in JJA is led by the decrease in surface wind speed in peak summer (Table S3), resulting from the decelerate of trade

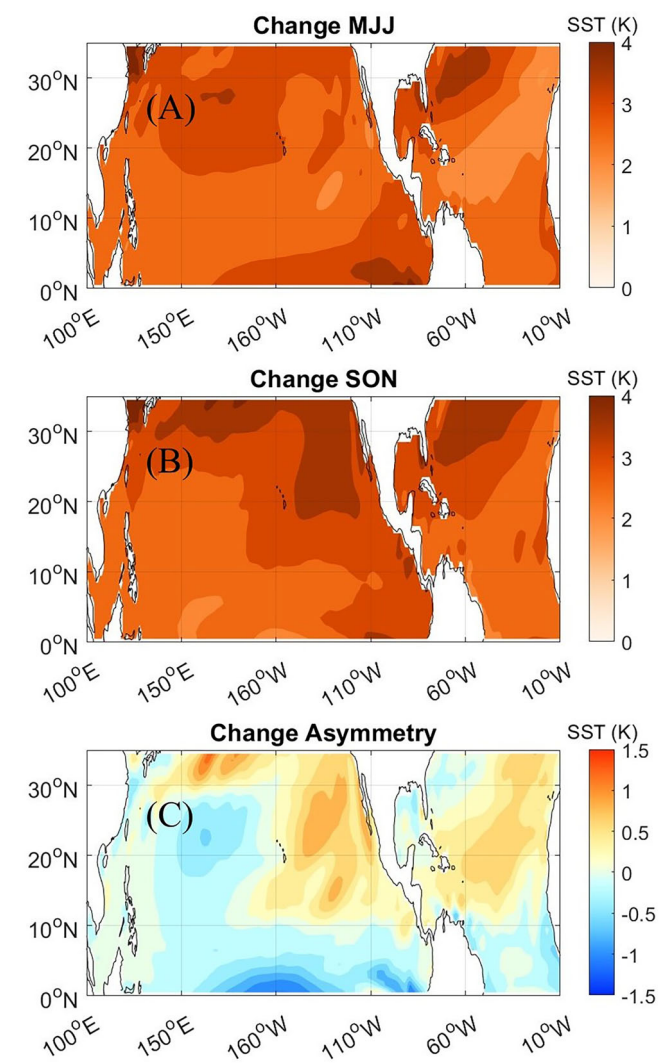


Fig. 6 | Change in SST from SPEAR simulations. Change in SST from SPEAR simulations in early season (May–July, A) and late season (September–November, B), and the temporal asymmetry of the changes (changes in late season minus changes in early season, C).

wind, which in turn is caused by the northward shift of the inter-tropical convergence zone in peak summer in ENA (Fig. S8).

Discussion

This study employed a high-resolution global climate model and found that under anthropogenic forcings of SSP5 8.5, the mean TC genesis date in ENP and NA is projected to shift to around 2 weeks later in the TC season by the end of this century. Such basin-wide shift in TC mean genesis date was not found in the WNP basin. Detailed analysis shows that for the WNP basin, late season shifts of TC genesis date are found in both South China Sea and the open WNP ocean north of 10°N, while an early season shift is found in the open WNP ocean closer to the equator. This study also links the shift in TC seasonal cycles in different sub-regions with large-scale environment and circulation and thermodynamic parameters (e.g., RH) changes. Table 2 summarizes the physical mechanisms for the TC seasonality changes for different sub-regions.

Recent studies^{10,16} reached different conclusions regarding the future shifts of TC season, while this study provided a more into-detail discussion on the spatial pattern and physical mechanisms of the changes using a coupled climate model with large ensembles. As shown in Table 2, for different sub-regions, different physical mechanisms, including the shifts in MT and Jet Stream, weakening of the zonal circulation in 10°N–25°N, as well as thermodynamic processes, can be influential on TC seasonality changes in different regions. The physical mechanisms found in this study, together with the change in ITCZ found in ref. 16, support the late-season shift of TC genesis for most of the sub-regions studied.

The late-season shift in TC genesis time for most of the sub-regions is somewhat different from that found in the other recent study on TC seasonality¹⁰. ref. 10 deduced a future early-season shift in the peak time of intense TCs by examining the favorable oceanic conditions for TC rapid intensification. Ref. 10 focused on TCs that reach at least category 4 (110 kts), while SPEAR underestimates intense TCs during 1979–2014 relative to historical observation, as most of the state-of-the-art global climate models do. To compare results of this study and ref. 10, here we define intense TCs as TCs that have the lifetime maximum intensity (LMI) higher than the 90th percentile of the LMI of all the SPEAR simulated TCs during 1979–2014, and examine the differences between the time that intense TCs reach their LMI during 1979–2014 and 2065–2100. The percentile selected corresponds to the threshold for Category 4 hurricanes in historical observations obtained using International Best Track Archive for Climate Stewardship best track data from 1979 to 2021. We found that even when we only

Table 2 | Summary of TC seasonality changes and the corresponding reasons

Sub-regions	Shift	Direct reason for GPI changes	Physical mechanisms
WNP I	Late	More mid-level humidification in late season	Shift in location of MT changes the location of deep convection, modulating the relative humidity.
WNP II	Slightly Late	More mid-level humidification in late season	Same as WNP I.
WNP III	Early	Decrease of low-level vorticity in late season	Shift in location of MT changes the location of the convergence zone, leading to low-level vorticity change.
ENP	Late	1 Decrease in wind shear in late season, while increase in wind shear in early season. 2 More mid-level humidifying in late season. 3 More increase in potential intensity in late season.	1 Equatorward (poleward) shift of Jet Stream in early (late) season, resulting in increase (decrease) wind shear in early (late) season. 2 Enhanced (weakened) zonal circulation in late (early) season resulting stronger (weaker) deep convection, which in turn results in more mid-level humidification in late season. 3 Larger SST increases in late season due to more energy storage during peak summer.
WNA	Late	1 Mid-level humidification (humidification) in late (early) season. 2 Decrease in wind shear in late season, while increase in wind shear in early season.	1 Stronger weakening of the zonal circulation in early season resulting in weaker convection in early season. 2 Equatorward (poleward) shift of Jet Stream in early (late) season, resulting in increase (decrease) wind shear in early (late) season.
ENA	Late	1 Increase (decrease) in potential intensity in late (early) season. 2 Decrease in wind shear in late season, while increase in wind shear in early season.	1 Larger SST increases in late season due to more energy storage during peak summer. 2 Equatorward (poleward) shift of Jet Stream in early (late) season, resulting in increase (decrease) of wind shear in early (late) season.

consider intense TCs for SPEAR, similar to ref. 10, the time for intense TCs to reach their peaks will be delayed by 5–10 days in ENP and NA (Fig. S9).

To enhance the reliability of the conclusion of late season shift of TC seasonal cycle found in one particular climate model (SPEAR), we also checked previous global statistical downscaling of TCs under SSP5 8.5 scenarios from six CMIP6 climate models (CanESM5, EC-Earth3, IP-SL-CM6A-LR, MIROC6, MPI-ESM1-2-HR, and MRI, methods detailed in the supplementary, ref. 6). The result (Figs. S8–S10) indicates that in ENP and NA, 5 out of 6 CMIP6 models show statistically significant late season shift of mean genesis date, while only 2 out of 6 CMIP6 models show statistically significant late season shift in WNP. It is evident that there are agreements among different climate models in projected signs of TC seasonal cycle shifts in ENP and NA, implying the late-season shift of TC seasonal cycle in these basins found by SPEAR can be supported by other climate models.

The delayed TC peak season close to coastlines may influence the proportion of TCs that can contribute to compound hazards in the future climate. However, the absolute changes in the compound hazards and impacts also depend on the changes in hazard frequency and intensity. For example, even though TC peak season will be delayed and thus less likely to overlap with the peak season of heatwaves, due to possible increase in heatwaves and TC intensity, compound TC-heatwave hazards may still increase, although the increase may be less compared to that without TC seasonal shift. If, for example, one assumes the TC seasonal cycle will not change, the increase of TC hazards in early season may be overestimated and the increase of TC hazards in late season may be underestimated if the intensity and/or frequency of TC landfall will increase in the future as in ref. 13. This bias in TC hazard estimation may influence the reliability of the projection of future TC-related compound hazards. In addition, a shift in TC landfall season could have significant socio-economic impacts. Relatively more TCs in the harvest season may impact agriculture and food security by influencing the production and supply chain of agricultural products, as was seen in 2005 after Hurricane Katrina²⁹. Relatively more TCs in the fall rather than in the summer can also interrupt the continuity of schooling in coastal areas that are impacted frequently by TCs³⁰. These compounding effects and socio-economic impacts due to the shift of TC seasonality should be further explored in future studies.

Materials and methods

SPEAR model and experiment design

The Geophysical Fluid Dynamics Laboratory Seamless System for Prediction and Earth System Research (SPEAR¹⁷) is used for the climate simulation. SPEAR consists of the AM4-LM4 atmosphere and land surface model, the MOM6 ocean model, and the SIS2 sea-ice model. In this study, the medium resolution configuration of SPEAR is used, where the horizontal resolution for the ocean/sea-ice model is $1^\circ \times 1^\circ$ and the horizontal resolution for the atmosphere is ~ 50 -km.

30 ensemble members are obtained using the SPEAR model. The detailed configuration of the All_forcing and Nat_forcing experiments can be found in Table S1 of ref. 31. In short, for the Nat_forcing experiment, also referred to as the control experiment, the radiative forcing is fixed at the 1921 level. For the All_forcing experiment, during 1921–2014 (in this study, we only use 1979–2014), the forcing is identical to time-varying historical natural and anthropogenic forcing, and during 2015–2100, the anthropogenic forcing is prescribed based on the Shared Socioeconomic Pathway 5.8.5³¹. Long-term pre-industrial climate simulations (over 3000 years) were generated by SPEAR, and their restart files were randomly selected to initialize the All_forcing and Nat_forcing experiments to form 30 ensemble members. Details about the experiment setup can be found in ref. 31.

Model simulated TCs were obtained from the 6-hourly outputs following ref. 31. The algorithm finds closed contours of sea level pressure anomalies and the 1-K temperature anomalies to identify the warm core and requires the detected storm to reach the relaxed wind speed criterion of 15.75 m/s given the 50-km horizontal resolution. The detected storms are required to have more than 36 h of lifetime to be counted in this study. The

genesis date of a TC is defined as the first time a TC reaches the above-mentioned criterion.

Genesis potential index (GPI) sensitivity analysis

The mathematical expression of GPI is shown in Eq. (1) in the main text. Due to its multiplication form, the changes in each parameter non-linearly interact with the changes in other parameters. To clearly separate the effect from each parameter, we define logGPI by taking the logarithm of Eq. (1):

$$\log \text{ GPI} = -2 \log(1.0 + 0.1 \times \text{SHR}) + 3 \log\left(\frac{RH}{50}\right) + 3 \log\left(\frac{Vp}{70}\right) + 1.5 \log(|VO \times 10^5|) \quad (2)$$

To perform the sensitivity analysis, we change the values of one term (with the coefficients) on the right-hand side for each grid in each month at a time from historical values (1979–2014) to SSP5 8.5 (2065–2100) values and calculate the change in log (GPI) solely due to this term. We then compute the difference in the change between late (SON) and early (MJJ) seasons (asymmetry).

Data availability

All the data and codes for generating the figures can be found in <https://doi.org/10.5281/zenodo.15354194>.

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Conceptualization: DX, HM. Methodology: DX, HM Investigation: DX Visualization: DX Supervision: HM, NL, MO Writing—original draft: DX Writing—review and editing: DX, HM, NL, MO.

Competing interests

All authors declare they have no competing interests.

Additional information

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