# Selected High Impact Journals

(截止到 2022 年 3 月)

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## 全文目录

**第一部分　作者简介**

**第二部分　文章选编**



黄刚，男，1971 年 11 月生于北京，理学博士，中国科学院首批特聘研究员（二级），中国科学院大学岗位教授，LASG 国家重点实验室季风与海气团队负责人，博士生导师，国家杰出青年基金获得者(2014)且优秀结题，国家中青年科技创新领军人才（2016），国家万人计划领军人才（2017）。1994 年本科毕业于成都气象学院，研究生师从于气象泰斗叶笃正院士，1999 年 9 月获得中国科学院大气物理研究所博士学位。目前担任多个国家自然基金项目以及多项国家任务的负责人，是基金委重点项目负责人，重大全球变化研究国家重大科学研究计划的课题组长等；并且担任中国气象学会动力气象委员会委员，大气科学常务编委，气象学报、气候与环境研究以及中国科学数据编委；同时担任 SCI 刊物 PeerJ, Frontiers in Earth Science 的编委以及 AAS 和 Advances in Meteorolog，Sustainability 的客座编委；担任 Nature 子刊以及 JC 等 42 种 SCI 刊物审稿工作；CNC-WCRP（中国气候研究委员会）委员以及中国科学院大学、南京信息工程大学、河海大学、成都信息工程大学等兼职教授以及博士或硕士生导师；也是香港中文大学研究生课程特邀授课教授等。曾获 2019 年度中国气象学会大气科学基础研究成果奖一等奖（排名第一），2017 年度广东省科学技术一等奖（排名第二），“学笃风正”优秀博士论文和科研论文奖,涂长望青年科技奖,全国优秀气象科技工作者,中国科学院“朱李月华”优秀教师奖（2012,2020），2012、2013 和 2018 年度中科院优秀研究生指导教师奖,并荣获 2013 年度赵九章优秀中青年科学奖以及 2014 年度卢嘉锡优秀导师奖等系列奖励。指导的博士获 2012、2013、2018 年度中科院优秀博士论文奖及其他青年奖项。截止到目前共发表 235 文章（包括 159 篇 SCI，62 篇 CSCD，14 篇书的章节），SCI 影响因子合计 853.779；第一作者（含通讯作者）118 篇（包括 88 篇 SCI，23 篇 CSCD，7 篇书的章节）；其中包括 Top 期刊 9 篇（Nature Climate Change 3 篇，Science Advances2 篇，Nature Geoscience3 篇, Annual Review of Earth and Planetary Sciences1 篇）。SCI 总引 6523 次，Google 学术引用 8583 次，H 指数 43，入选美国斯坦福大学发布了"2020 全球前 2%顶尖科学家榜单"，气候变化研究领域全球最具影响力的 1000 位科学家名单以及 ESI 全球地球科学领域科学家排名前 1%高引用科学家目录，并且拥有 3 项计算机软著（第一完成人）和 1 项专利（第一完成人）。

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# 黄刚高影响文章选编

## 目　录


1. **Huang, G***, Z. Xu, X. Qu, J. Cao, S. Long, K. Yang, H. Hou, Y. Wang and X. Ma, 2022: Critical climate issues towards carbon neutrality targets, Fundamental Research , https://doi.org/10.1016/j.fmre.2022.02.011
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Perspective

# Critical climate issues toward carbon neutrality targets

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## 1. Introduction

On September 22, 2020, Chinese President Xi Jin-Ping solemnly announced that "China will scale up its Intended Nationally Determined Contributions by adopting more vigorous policies and measures, aiming to have $CO_2$ emissions peak before 2030 and achieve carbon neutrality before 2060" at the General Debate of the 74$^{th}$ Session of the United Nations General Assembly. The so-called carbon neutrality means a state of net-zero carbon dioxide emissions. This ritual commitment shows the responsibility of China as a global superpower to follow a "green, low-carbon, high-quality" development path. Our series of deployments to address climate change and carbon neutrality will have far-reaching impacts on the energy structure, industrial structure, and socioeconomic development of China and even the world. It will also have considerable effects on technological progress and global governance. The ultimate goal of carbon neutrality is to unify international countries to jointly achieve the temperature target of the Paris Agreement, i.e., to control the global average temperature rise within 2 ˚C and to strive for only 1.5 ˚C above preindustrial levels. Different from the traditional carbon emission scenarios [1], the Paris Agreement established the first scenario in which temperature is regarded as a target to constrain climate change, pushing a new direction of climate research. Achievement of the target would require a decrease in atmospheric $CO_2$ concentration [2,3]. However, the World Climate Research Programme (WCRP) stated in 2018 that "the scientific community lacks consensus on the climate effects of $CO_2$ removal" and launched the Carbon Dioxide Removal Model Comparison Program [4]; as of 2021, the data were still renewing, and relevant studies were few.

Climate change is closely related to human society. Current global warming, mainly caused by the massive emissions of greenhouse gases, especially carbon dioxide, has dramatically affected human production and life. Despite a continuous increase in $CO_2$ since the 20$^{th}$ century, the global average temperature has shown a stepwise increase, including warming in the early 20$^{th}$ century, cooling in the mid-20$^{th}$ century, rapid warming in the late 20$^{th}$ century, and a hiatus in the early 21$^{st}$ century [5–7], underpinned by the nonmonotonic evolution of surface temperature [8,9]. During the period of warming, global temperature change followed the familiar "temperature–$CO_2$" relationship, but during the hiatus period, i.e., the early 21$^{st}$ century, the relationship was decoupled. This implies that the relationship between $CO_2$ concentration and surface temperature, which is of most concern to humans, is not linear and may involve complex dynamic feedbacks. Previous studies have suggested that the coupled and decoupled "temperature–$CO_2$" relationship can be due to the different time-scale responses of the climate system to anthropogenic activities, such as a slow response due to the large heat capacity of the ocean to anthropogenic radiative forcing [10–13]. On the other hand, the internal variability of the climate system, such as the cold phase of the Pacific Decadal Oscillation (PDO), may offset the warming effect of human activities [5,6]. In the future, under the scenario of "emission peak" and "carbon neutrality," our ultimate goal is to limit the rise of global surface temperature. Our control of carbon emissions will finally act on climate change. The management of $CO_2$ concentration is only a method, not the ultimate goal. Therefore, a scientific understanding of the "temperature–$CO_2$" relationship (especially around some key turning points) is essential for planning and coping with climate change in the coming decades.

2020 was an extraordinary year with worldwide climate extremes. The sudden epidemic of COVID-19 early in the year led to a severe reduction in our productivity. As a vast industrial country, the stagnation in productivity caused a significant decrease in $CO_2$ and aerosol emissions. With effective epidemic prevention measures, our productivity levels began to recover at the end of the year and were coming on strong in 2021, while the emissions in the rest of the world continued to decline. COVID-19 forced us to passively experience an emission reduction. Reductions in $CO_2$ or aerosols may cause an asynchronous response of the ocean and atmosphere in a short period, as the response of the ocean to radiative change is slower than that of the land and atmosphere, which may cause potential ecological, environmental, and engineering safety risks. Extreme weather tends to become more frequent under global warming; for example, the record-breaking Yangtze River floods in China triggered

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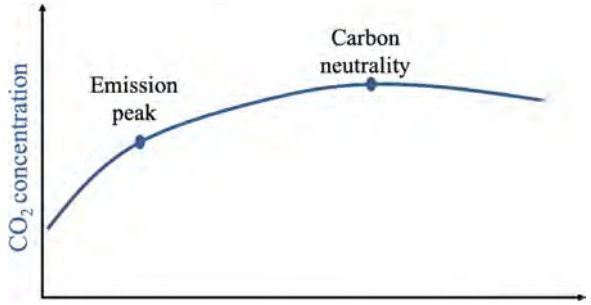



**Fig. 1.** Schematic diagram of the evolution of $CO_2$ concentration in a carbon-neutral background.

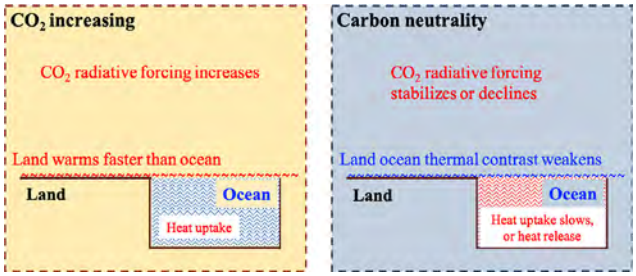



**Fig. 2.** Diagrams of climate-dynamical processes at different stages of atmospheric $CO_2$ concentration change. (a) During rising stages, $CO_2$ radiative forcing is the main driving force of the climate system. Due to the small heat capacity, land warms rapidly with the increasing of $CO_2$ radiative forcing; the ocean's response is slow owing to its large heat capacity. The thermal difference between land and ocean increases, which enhances monsoon circulation. (b) After carbon neutrality, $CO_2$ radiative forcing stabilizes or decreases, while the ocean is still slowly absorbing heat and warming. The land-ocean thermal contrast diminishes, weakening the monsoon circulation; if the $CO_2$ concentration decreases dramatically, the deep ocean will release heat upward to decelerate the cooling of the sea and land surface.

by the strong anomalous Mei-yu event in 2020; the dust storms in northern China and thunderstorms and windy weather in southern China in early 2021; the daily rainfall of more than 100 mm in South Xinjiang in June; and the once-in-a-thousand-year event in Zhengzhou on July 20, of which the hourly rainfall exceeded 200 mm. Beyond China, these extreme weather events include the wildfires in Australia and California in 2020, the snowstorm on the eastern coast of Canada, the heatwave in India and France, the extreme cold weather in Texas in February 2021 that led to widespread, long-term power outages, the extreme precipitation in Europe in spring, etc. These extreme events resulted in tremendous damage to the production and lives of human beings. Should these severe weather and climate disasters be attributed mainly to human activities or natural variability? The actual physical processes of the "emission peak and carbon neutrality" scenario may be more complex. A better comprehension of the changes in the proportion of the climate disaster process is of high scientific and application value for predicting the intensity, area, and duration of extreme weather and climate disasters under future scenarios.

The target of carbon neutrality is an integrated and extremely complex issue that is widely associated with climate, ecology, energy, environment, society, etc. It results from interactions of multiple systems, such as land, atmosphere, ocean, and cryosphere. At present, the climate impact of carbon neutrality is still unclear, and it may be related to planning a reasonable carbon emission path to realize the optimal layout of energy, industry, the ecological environment, and other fields to achieve carbon neutrality with relatively small socioeconomic costs. To date, there are several questions that still need to be answered. How will the regional climate over China and the global climate respond to the three critical stages of the carbon neutrality target (Fig. 1), i.e., from now to the "emission peak", from the "emission peak" to "carbon neutrality," and from "carbon neutrality" to the end of this century? What will be our stance on achieving an "emission peak"? Are there feedbacks from climate change on carbon emissions or uptakes in both directions? What are the possible feedbacks? How much uncertainty is there? To address these questions, it is essential to understand the relationship between atmospheric $CO_2$ and temperature, i.e., to address the climate under carbon neutrality, which is now a critical task.

## 2. Understanding climate dynamics under carbon neutrality

During the historical period, researchers have conducted climate analysis with instrumental data, which, however, are unavailable in the future carbon-neutral scenario. Earth system models and climate dynamics theory are needed to comprehend and project climate change.

Under a carbon-neutral scenario, accurate prediction of climate–carbon feedbacks and the hazards of weather and climate highly depends on the precise projection of the future climate. The scenario of carbon neutrality will be significantly distinguished from the scenario we have experienced since the industrial revolution. Additionally, the corresponding climate dynamics are different from the well-known stages during which $CO_2$ is rising. Before neutrality, the $CO_2$ concentration will continue to increase, and radiative forcing during this period will be the leading factor in global warming, dominated by the rapid responses of the climate system (especially land and mixed-layer oceans). The heating effect due to the greenhouse gas (GHG) effect notably (approximately 90%) enters the ocean, which acts as a "heat reservoir" [1,14]. After carbon neutrality, GHGs will stabilize or decline. The slow responses of the climate system (the slow absorption or release of heat from the deep ocean) will gradually increase. The "heat reservoir" effect may resist the cooling effect led by the decline in $CO_2$ concentration or slowly warm the global surface when the $CO_2$ concentration stabilizes. Supposing that the $CO_2$ concentration decreases too rapidly, the ocean may even overwhelm GHGs and become the driver of global climate change, as shown in Fig. 2.

China features a typical monsoon climate. The thermal contrast between land and ocean plays an essential role in forming the East Asian monsoon [15]. Before carbon neutralization, the land warms more rapidly than the ocean, strengthening the thermal contrast and favoring the continued enhancement of the East Asian summer monsoon. After carbon neutralization, the ocean may still warm or cool slowly due to its thermal inertia, leading to a weaker thermal contrast and weakening the East Asian summer monsoon [16,17]. All of the above factors may have impacts on the climate in China.

Multiple studies have suggested that the El Niño–Southern Oscillation (ENSO) amplitude will strengthen, and extreme El Niño events will continue to increase during the carbon-neutral stage [18,19]. During extreme El Niño events (e.g., the 2015–2016 super El Niño event), the carbon emissions from the biosphere will increase due to high temperatures and droughts in the tropics [20]. The response in the oceans under global warming may alter the uptake and storage of $CO_2$, which can impact the realization of carbon neutrality. For example, the North Atlantic Ocean is one of the major global carbon sinks. Due to deep convection in the sinking branch of the AMOC, a large amount of carbon is sequestered in the deep North Atlantic; therefore, a weakening of the AMOC will reduce its carbon storage capacity [21], thus affecting the atmospheric carbon concentration. In addition, under the low-carbon emissions scenario, the North Atlantic Ocean would potentially gradually replace the Southern Ocean as the focus of global ocean heat uptake due to the weakening of the Atlantic Meridional Overturning Circulation (AMOC) [22]. Under the carbon neutrality goal, future anthropogenic aerosol emissions will continue to decrease. The subsequent warming effect will be superimposed on the cooling effect caused by $CO_2$ reduction, resulting in an increase in surface warming and a decrease in the rate of temperature decline, thus prolonging the time needed to achieve

the Paris Agreement's low-temperature rise target [23]. Therefore, when designing specific pathways to achieve carbon neutrality, the critical climate impacts of anthropogenic aerosols need to be considered.

Although the scientific community has some preliminary understanding of the above processes, some difficulties appear in research:

(1) Uncertainty in atmospheric $CO_2$ concentrations in the future. The future pathway of $CO_2$ concentration is a result of net carbon emissions from countries all over the world. At present, although most countries have proposed related carbon neutral policies, there is still much uncertainty regarding the pathway of $CO_2$ emissions as well as its concentration. This increases the difficulty of climate prediction.
(2) Uncertainty of numerical simulation. The Earth system model is an ideal tool to carry out climate prediction in the carbon neutrality scenario. Many physical processes in the Earth system model are simplified, and the details of weather and climate processes are not fully described, which introduce simulation biases. Moreover, the lack of observation data and knowledge of some regional climate phenomena makes it difficult to judge whether the presentation in the Earth system model is reasonable. This uncertainty may affect other regional climate phenomena.
(3) The influence of internal climate variability. In addition to human activities, the climate is also influenced by internal climate variability. For example, during the first decade of the 21st century, atmospheric $CO_2$ continued to rise, but global temperature displayed a hiatus due to the cooling of the eastern Pacific Ocean [5]. Considering that in numerical simulation experiments, internal climate variability is stochastic, it is a challenge to force out the variability. This makes climate prediction more difficult.

In general, the scientific understanding of climate dynamics under carbon-neutral scenarios is insufficient. The impact of these dynamic processes on East Asia and the global climate is less understood. In addition, the precise quantification of the climate–carbon feedbacks under future climate relies on a deeper understanding of climate dynamics and a more accurate prediction of climate change under a carbon-neutral scenario. Therefore, we need to conduct independent research and systematically understand the climate dynamic processes under carbon neutrality scenarios, as well as their impacts on climate change in China and the globe. A deep understanding of these issues may enable us to accurately and scientifically project the risk of climate in China and the world under the influence of carbon-neutral policies, the climate conditions in critical areas of carbon sources and sinks, and the climate–carbon feedback before and after carbon neutrality. It provides scientific support to the rational layout of energy, industry, ecology, and the environment in China and makes carbon neutrality better and more stable.

## 3. Quantitative analyses of climate–carbon feedback

At present, while there has been some relevant research on carbon–climate feedback, there is still considerable uncertainty in the climate–carbon-cycle model results that still need to be further developed [24–26]. In the field of climatology, recent studies have primarily focused on one-way impacts, i.e., carbon emissions on climate change, while in fact, the carbon cycle and climate are interactive. The IPCC Fifth Assessment Report (IPCC AR5) concludes that there is positive feedback between the carbon cycle and climate: under the future scenario, global warming may weaken the carbon sink of ecosystems, leading to higher $CO_2$ concentrations in the atmosphere and further reinforcing global warming. The above understanding provides some implications: (1) If we ignore the weakening of the carbon sink under global warming, the timing of carbon neutrality achievement may be delayed. (2) If the atmospheric $CO_2$ concentration does not decrease enough after carbon neutrality, the global temperature may rise slowly and continue to weaken the global carbon sink. Therefore, all countries cannot evade the efforts to control carbon emissions.

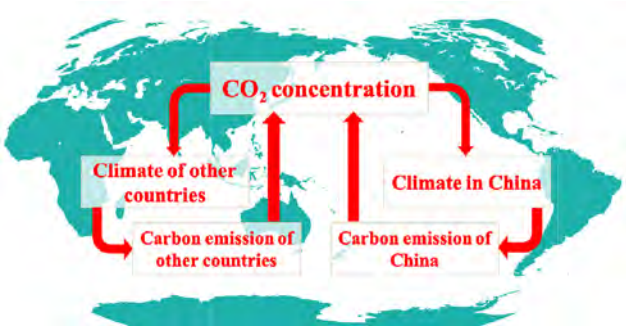



**Fig. 3.** Relationship between China and other countries in terms of net carbon emissions and climate.

However, there is uncertainty in understanding some key parameters in the climate–carbon feedback (e.g., feedback strength, critical thresholds, and spatial distribution), which significantly constrains the planning of carbon-emission reduction. Supposing that we do not have an accurate understanding of the climate–carbon feedback; then we may fail to achieve the optimal deployment of the development of energy, industry, ecological environment and other fields in China. It may bring additional socioeconomic losses. Therefore, there is an urgent need to conduct research on the carbon cycle and climate feedback. In addition, in recent years, the WCRP updated the "Grand Challenges" on climate research, in which carbon–climate feedback was listed as one of the seven challenges. This implies the importance of climate–carbon feedback.

## 4. Interaction between China/global climate and carbon emissions

$CO_2$ can mix in the atmosphere and affect the climate through changes in its concentration on a global scale. That is, the climate change of each country is not solely determined by the net carbon emissions within the territory of each country but depends on the net carbon emissions on Earth. Therefore, on the issue of climate change and carbon emissions, all countries on the worldwide scale are not isolated but are closely linked communities [1,27-29], as shown in Fig. 3.

China has developed a long-term plan to achieve carbon neutrality, which will significantly contribute to global climate mitigation and carbon reduction. Based on the understanding of carbon–climate feedbacks and climate dynamics under the "carbon neutrality" scenario, this makes it possible to estimate the contribution of China's carbon neutrality policy to global warming mitigation, climate disasters, and ecological carbon sinks. On the other hand, other countries might also influence the domestic carbon emissions in China. Imagine an extreme scenario under which our country achieves carbon neutrality while the net $CO_2$ emissions of other countries are positive. Then, the $CO_2$ concentration still rises, leading to a higher-than-expected temperature increase in our country, further weakening the (ecosystem) carbon sink and destroying the carbon-neutral state of China. Considering this, we can also rely on the further study of climate–carbon feedback and climate dynamics to assess the impact of carbon emissions from other countries on our carbon-neutral status. These two aspects are important issues to be considered in climate diplomacy.

Based on an in-depth understanding of the carbon–climate feedback and carbon-neutral climate dynamics, it is feasible to conduct a study on the interaction among China's carbon neutrality, global climate, and carbon emissions with the aid of sophisticated Earth system models. Conducting research in this area and quantifying the contribution of China's carbon neutral policies to global climate and carbon emissions helps increase China's discourse power in climate diplomacy. Moreover, assessing the impacts of carbon emissions of other countries on climate

and carbon emissions in China and timely adjusting China's deployment in the fields of energy, industry, ecology, and the environment prudently favor the achievement of carbon-neutral goals.

## 5. Solutions to scientific problems associated with carbon neutrality

The major political plan of carbon neutrality gives rise to a series of scientific problems. Future human-induced GHG change is a physical problem and a complex problem involving politics, society, and the economy. Anthropogenic perturbations can result in impacts and consequences that threaten the security of the Earth system. Due to the complexity of the Earth system and the nature of nonlinear responses, changes in the relationship between humans and nature in the short term will inevitably cause significant uncertainty in the prediction of the future environment. Thus, it is worthwhile to start from the following three aspects.

(1) One solution is climate dynamic research and Earth system model development. The understanding of climate dynamics under carbon neutrality relies on the theoretical study of climate dynamics and Earth system models. The Earth system model is the only dynamic tool that we use to predict the future. Climate dynamics studies can be used to validate Earth system models and jointly contribute to accurate estimates of climate–carbon feedbacks together with models. With a proper understanding of carbon-neutral climate dynamics and the key quantitative indicators of the climate–carbon feedback, we can use them to assess the interactions between China and the globe on climate and carbon emissions.
(2) The second solution is to develop multidisciplinary theoretical research, emphasizing the interactions of multiple systems of the Earth, including the atmosphere–ocean, atmosphere–land, and other multilayer interactions, as well as the two-way feedback of climate and carbon. Carbon neutrality is not only a scientific issue but also a complex sociological topic. It is important to focus on the intersection of social sciences (e.g., economics and international relations) with climatology. Currently, the promotion of interdisciplinary research is an inevitable trend in the development of Earth sciences. The understanding of carbon neutrality is also accompanied by the rapid development of computer technology, which provides computing power for large numerical calculations.
(3) The third solution is to effectively gather human resources, extensively recruit outstanding scientists in related fields, and bring into research teamwork synergy. For example, we could build large-scale computer devices to provide strong support for large-scale scientific calculations and establish carbon-neutral research centers. In particular, attention and emphasis should be placed on institutional arrangements and national scientific policies to strengthen the scientific innovation of China. For example, funding for fundamental research should be strengthened, and young researchers, especially scholars who have recently received their doctoral degrees, should be given more opportunities to explore independently, and the evaluation mechanism should not be paper-centered but should be more varied.

## 6. Conclusion and prospects

In the process of energy change and carbon neutrality emerging globally, greenhouse gases such as $CO_2$ in the atmosphere will be influenced by significant anthropogenic perturbations. Although low-carbon emissions may slow atmospheric warming, the overall trend of climate warming will continue for a long time. Achieving carbon neutrality is likely to lead to changes in global and national weather hazard patterns and potentially decrease the uncertainty of climate variability and the occurrence of climate extremes, with important implications for global security. Therefore, the feedback and response of the Earth system to anthropogenic $CO_2$ emissions after carbon neutrality and its impacts will be scientific issues worthy of attention in the future. Carbon neutrality is closely related to domestic production and lives, and there is little time left for planning. This set of factors has given rise to key scientific questions that need to be addressed in the context of carbon neutrality, mainly including the following:

(1) Carbon-neutral climate dynamics research (understanding the key climate dynamic processes under carbon-neutral scenarios, including the fast and slow responses of the climate system, internal climate variability, and their impacts on our domestic and global climate).
(2) Two-way feedbacks between the regional and global climate–carbon feedback (to understand the spatial distributions of the crucial quantitative indicators, such as feedback coefficients and critical thresholds and climate–carbon feedbacks in China and worldwide).
(3) The contribution of China's carbon neutrality to global climate and carbon emissions (assessing the contribution of China's carbon neutrality policy to global warming mitigation, climate disasters, ecological carbon sinks, etc. and evaluating the impact of other countries' carbon emissions on China's carbon neutrality status and route).

In summary, climate change has profoundly affected human production and life. Carbon neutrality is not a purely scientific topic but a complex and comprehensive issue. Established policies and emission reduction pathways might be highly affected by unexpected events, such as the frequent occurrence of extreme weather disasters and outbreaks of epidemics in recent years, which make the "carbon peak and carbon-neutral" scenario face great uncertainty. There is enormous uncertainty but also a massive challenge to the scientific community. However, we should believe that we will make a breakthrough in climate issues in the context of carbon neutrality with our joint efforts, which will serve our country's carbon emission policy at different stages.

## Declaration of competing interest

The authors declare that they have no conflicts of interest in this work.

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News & Views

# The first 5-year clean air action did increase the blue days in winter over Beijing-Tianjin-Hebei

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China has taken great efforts to mitigate its long-standing environmental problem in recent years. During the first 5-year Clean Air Action (2013–2017), remarkable reductions of primary pollutants such as $SO_2$ and fine particle matter ($PM_{2.5}$) have been achieved in China, especially in Beijing-Tianjin-Hebei (BTH) region [1,2]. After the explosive increase of concern on the Beijing Blue Events in 2015, the direct changes of blue days in China have received widespread attention [3,4]. However, less air pollution does not directly correspond to more blue skies. For example, days with high cloud cover or precipitation do not belong to blue days even there is less air pollution. In addition, inconsistency between dropping PM, unchanged low visibility events, and a remarkable increase of aerosol optical depth can be found even during the COVID-19 lockdown period [5–7]. Beyond that, unfavorable meteorological conditions counteract the effects of reduced emissions, resulting in frequent haze in the BTH region [8,9]. Therefore, there still lacks a direct demonstration of more blue days due to emission cuts.

Along with living standard enhancement, there is an advanced understanding that the blue day should be a day with blue sky and great air quality. Then a novel index named the Chinese Blue Days Index (CBDI) has been put forward to reveal the long-term spatiotemporal change of Chinese blue days [10]. However, the CBDI represents a national average level, which is inappropriate for BTH. In addition, earlier works that have focused on daily observational data are unable to rule out the effects of meteorology and air pollution.

To quantitatively distinguish the effect of aerosol on blue days from the meteorology, the effectiveness of the first 5-year Clean Air Action is assessed by using the Weather Research and Forecasting Model coupled with chemistry (WRF-Chem) in this study. Two sensitivity experiments were designed to investigate the effects of emission control strategies since 2013. The baseline experiment is the control simulation, which is conducted by anthropogenic emissions in 2017 and fully coupling of aerosols and meteorology. The sensitivity experiment has the same configurations apart from anthropogenic emissions in 2012. Our experiments are conducted for two typical winters as the blue days in winter are most concerned. The first one is the last winter of the first 5-years Clean Air Action (i.e., Nov., Dec. 2017, and Jan. 2018), and the second one is the first winter after the first 5-years Clean Air Action (i.e., Nov. and Dec. 2018). Due to a lack of meteorological observations for model evaluation, our experiments do not include Jan. 2019. To focus on the BTH region, a triple nesting scheme was adopted (Fig. S1 online). Model configuration can be found in Table S1 (online). Detailed descriptions of the materials, methods, and model set are summarized in Supplementary materials.

The BTH blue days (BBDs) are firstly defined with three criteria following methods of Wang et al. [10]: no precipitation (Non-Prep), low cloud cover (LCC) $\leq$ 30%, and visibility $\leq$ 15 km at 14:00 BJT. The model performances are evaluated with *in situ* observations in BTH. The simulated temporal variations of visibility, LCC, and Non-Prep match well with the observed ones, with correlation coefficient values higher than 0.88, 0.76, and 0.84 respectively (Fig. S2a–c online). Consequently, the simulated BBDs agree well with the observed ones, with a correlation coefficient beyond 0.85 and a high accuracy beyond 80.23% (Fig. S2d online). These indicate a good model performance. The spatial distributions of the simulated visibility, LCC, Non-Prep, and BBD show patterns similar with the observed ones (Fig. S3 online). These results indicate the rationality of evaluating BBDs changes under different emission scenarios.

A widespread decline in anthropogenic emissions during 2013–2017 has been documented in many studies [1,11,12], while few have focused on the variations in winter. Based on the Multi-resolution Emission Inventory for China (MEIC v1.3; http://meicmodel.org) [13], the average emissions of primary aerosol species and trace gases during winter 2012 and 2017 in BTH are shown in Table S2 (online). Compared with 2012, the remarkable effectiveness of Clean Air Action can be detected for $SO_2$ and PM coarse (coarse mode of particle matter, calculated by the difference

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between $PM_{10}$ and $PM_{2.5}$), whose amplitudes are reduced by more than 50% in winter for BTH. By contrast, minor declines are detected for $NO_x$ and $NH_3$, which are below −26% and −15% respectively. These results are consistent with the earlier studies [11,12,14]. Nonetheless, the emissions of volatile organic compounds (VOCs) still increase beyond 12.66% and 16.84% in Beijing and Tianjin and display a weak reduction below −6.58% in Hebei. These indicate that more attentions should be paid to the VOCs, $NO_x$, and $NH_3$ emission controls in the future.

Accordingly, a noticeable cut-emission effect on blue days is manifested in Fig. 1a. There is a comprehensive increase of BBDs in BTH and surrounding areas compared with that using emission in 2012, which is closely related to the cut-aerosol distribution (Fig. S4a–f online). In comparison, southeastern Beijing (over 45%), southern Tianjin (over 65%), and Hebei (over 65%) show the highest increasing rate, with a maximum exceeding 53 d at an increasing rate of 157.14%. We also notice a mismatch between the most apparent emission reduction in Beijing and the most obvious BBDs increase in southern Hebei (Fig. S5a online). This may be caused by the variations of VOCs and the corresponding climate-aerosol feedback effect, which needs further study.

Furthermore, the major contributor to the increase of BBDs is detected. With fixed meteorology, the simulation suggests that reduction of emissions would lead to significantly higher visibility (Fig. 1b, passing the 95% confidence level). In addition, the most notable increase of visibility (over 46%) runs slightly from the northeast to southwest, which is consistent with the BBDs variations. Overall, the correlation coefficient between visibility and BBDs is beyond 0.93. It should be mentioned that winter is the "dry" season for BTH with low LCC and little precipitation, therefore the difference of LCC and Non_Prep between two MEIC emissions can not pass the 95% confidence level (Fig. S6 online) [15]. This demonstrates the dominant role of visibility and the minor effects of LCC and precipitation on BBD increase. Meanwhile, it emphasizes the significance of emission control in the increase of BBDs in recent years.

The variations of regional hourly average BBDs and visibility under different emissions in BTH are plotted in Fig. 1c. Clearly, emission controls contribute to greater visibility, with a mean increment of 2.57 km. Meanwhile, it helps to weaken visibility decline during serious pollution episodes (in blue bars in Fig. 1c). Meantime, it is emphasized that emission reduction brings into a

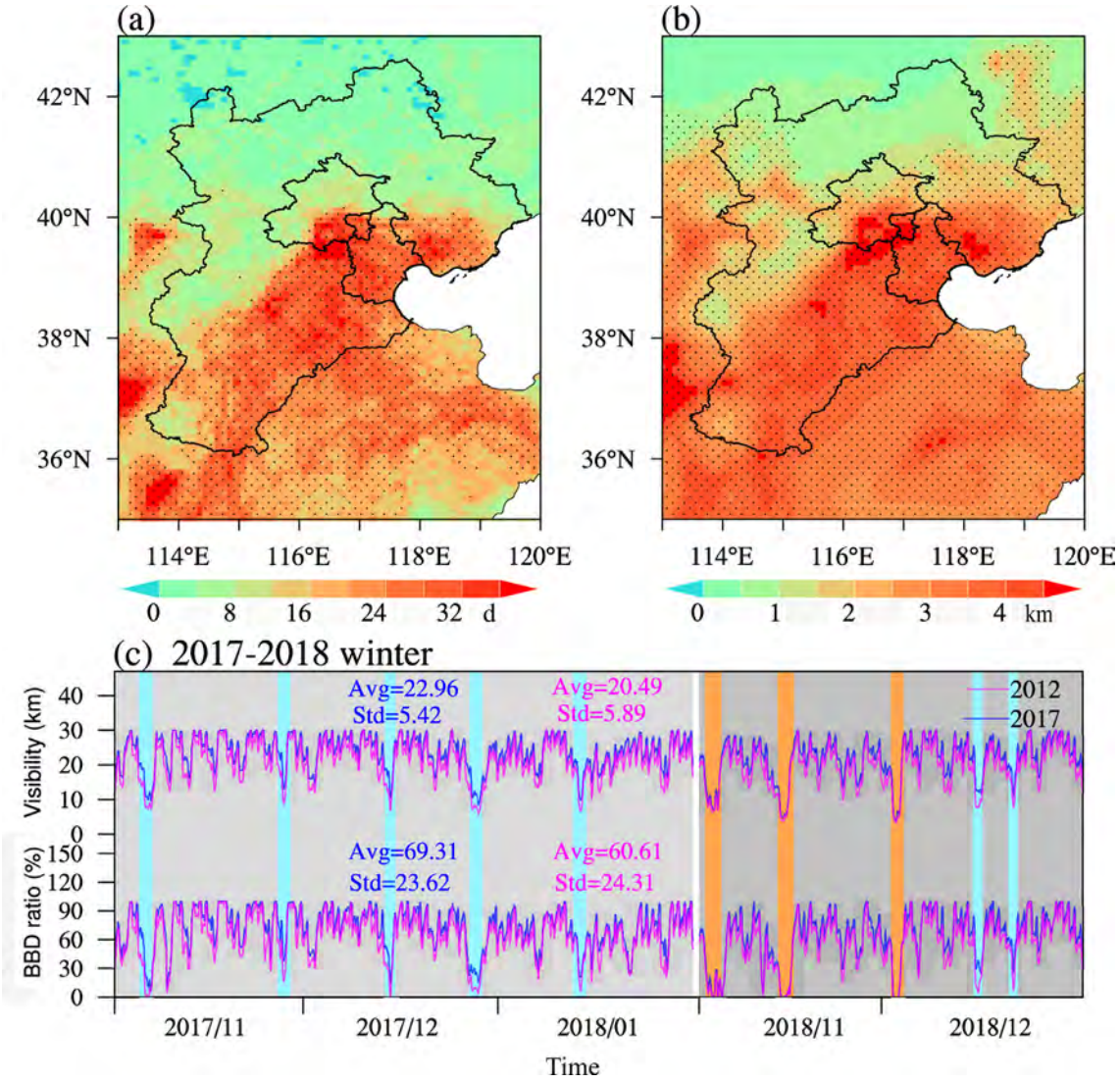



**Fig. 1.** Spatial distributions of differences of simulated total blue days (Sum_BBD, unit: d) (a), and average visibility (unit: km) (b) under different MEIC (2017 and 2012) during winter 2017 and winter 2018 over BTH. The black dots mean the changes pass the 95% confidence level by *t*-test. (c) The time series of the regional hourly average BBDs occurrence rates (BBD ration) over BTH. The blue bars represent the serious pollution episodes have been improved by emission control measures, and the orange bars mean the opposite.

better-than-better effect, meaning that control measures improve the visibility during relatively higher-visibility days efficiently, such as Nov. 16–21, 2017, and Dec. 30–31, 2017. By reducing the amplitude of oscillation at an hourly scale, lower standard deviations can be detected for visibility. Following the variations of visibility, the emission controls seem to be effective to improve most of the extreme pollution events and favor a bluer-than-bluer mechanism during relatively bluer days on the regional scale in the same period. We also notice that the extreme pollution episodes in Nov. 2018 and early Dec. 2018 seem to be a refutation for strict emission measures (in orange bars in Fig. 1c). This suggests that the unfavorable influence of meteorology may play a more important role for regional non-blue days in BTH on that days.

On average, the regional average BBDs occurrence rates over the 16 sites have risen from 54% to 65%, which is consistent with the regional mean results (from 60% to 69%). Regionally, those control measures remarkably increase the Beijing blue days from 22.32 days/month (d/m) in 2012 to 25.73 d/m in 2018, from 12.4 to 17.67 d/m in Tianjin, and 18.91 to 22.01 d/m in Hebei, respectively. Overall, a regional mean BBDs increase beyond 2.79 d/m is revealed. It is also notable that the emission-cut measures lead to a weaker correlation coefficient between visibility and BBDs in BTH. The correlation coefficients decrease from 0.97/0.93/0.93 to 0.94/0.92/0.91 in Beijing/Tianjin/Hebei, respectively. This means that the dominant role of visibility on BBDs may be diminished with its improvement, which needs further study.

The impacts of Clean Air Action since 2013 on the blue days in winter over BTH are explored in this study. Reduction of emissions leads to significantly higher visibility and widespread increase of BBDs in BTH by an average of 2.79 d/m and a maximum exceeding 53 d. Although the unique emission controls may favor more extensive blue days and reduce extreme pollution episodes in China, more efforts should be implemented to achieve more blue days effectively in the future, which will provide a reference to future policymaking from a new perspective. Further study should be conducted to qualify the positive feedback mechanism of emission-cut and blue days and the affections of the newest Blue Sky Protection Campaign Plan should be evaluated.

## Conflict of interest

The authors declare that they have no conflict of interest.

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## Appendix A. Supplementary materials

Supplementary materials to this article can be found online at https://doi.org/10.1016/j.scib.2022.01.009.

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ARTICLES

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# Sulfur emissions from consumption by developed and developing countries produce comparable climate impacts

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**Regional consumption activities supported by domestic production and international trade have led to substantial aerosol-related emissions worldwide. Here we quantify sulfur dioxide emissions associated with consumption by developed and developing countries and assess the resulting climate impacts using an Earth system model. We find that although the consumption-associated emissions of developed countries are 40% less than those of developing countries, they lead to similar impacts on global mean surface air temperature and precipitation. This is because the effective radiative forcing induced per emission is greater for developed countries, which we attribute to the emissions being located at higher northern latitudes and being more evenly distributed zonally. Emissions from developing countries have a greater impact on temperature and precipitation over the tropical monsoon regions of China and India. Our results demonstrate the importance of trade and emission region in determining how consumption translates into global climate impact.**

Climate Action and Responsible Consumption and Production are two closely linked Sustainable Development Goals of the United Nations. Consumption activities affect the climate through greenhouse gases and aerosols associated with production to supply consumption. Raw materials extracted to meet global final consumption demands reached 92 billion metric tons in 2017 and will probably increase to 190 billion metric tons by 2060 without concerted action[1], and extraction, processing, use and disposal of these materials often lead to substantial emissions. Many emission pathways of 1.5 °C warming require shifts towards sustainable consumption[2]. As international trade allows consumption of any region to be supplied by production worldwide, this affects the magnitude and geographical pattern of consumption-associated emissions[3–5]. Understanding the climate response to consumption-associated emissions offers crucial information to support policymaking and international cooperation in climate change mitigation action from a consumption perspective.

For aerosols, which are short-lived, the spatial distribution of consumption-associated emissions determines the distribution of their atmospheric loadings and thus exerts spatially inhomogeneous forcing[6,7] to the nonlinear climate system. Emerging evidence has suggested strong nonlinearity in the climate response to the magnitude[8,9] and location[10–13] of aerosols. For example, a given amount of aerosol emissions released in Western Europe may cause 14 times as much global average cooling as when emitted in India[11]. These studies are based solely on idealized model experiments by scaling the emissions with an arbitrary factor or by putting the same amount of emissions to different locations. Many studies have analysed the climate impacts of aerosols associated with each region's own production activities and resulting (realistic) emissions[14–17]. The climate response to aerosols associated with the same region's consumption could differ substantially due to the large differences in the magnitude and spatial pattern of emissions and forcing. Yet this consumption-induced climate response has never been assessed explicitly. One recent study has estimated the direct radiative forcing of aerosols embedded in trade[6], but it does not quantify how such forcing affects temperature and precipitation.

In this article, we provide an assessment of the effects of aerosols from consumption of developed and developing countries on global temperature and precipitation. We quantify the climate response to sulfate aerosol forcing due to consumption-associated emissions of sulfur dioxide ($SO_2$) in 2014, the latest year for which all data and models are available. For this purpose, we combine an emission inventory[18], a multiregional input–output table[19] and the fully coupled Community Earth System Model version 2 (CESM2)[20] (Methods). Identification of developing (including emerging markets) and

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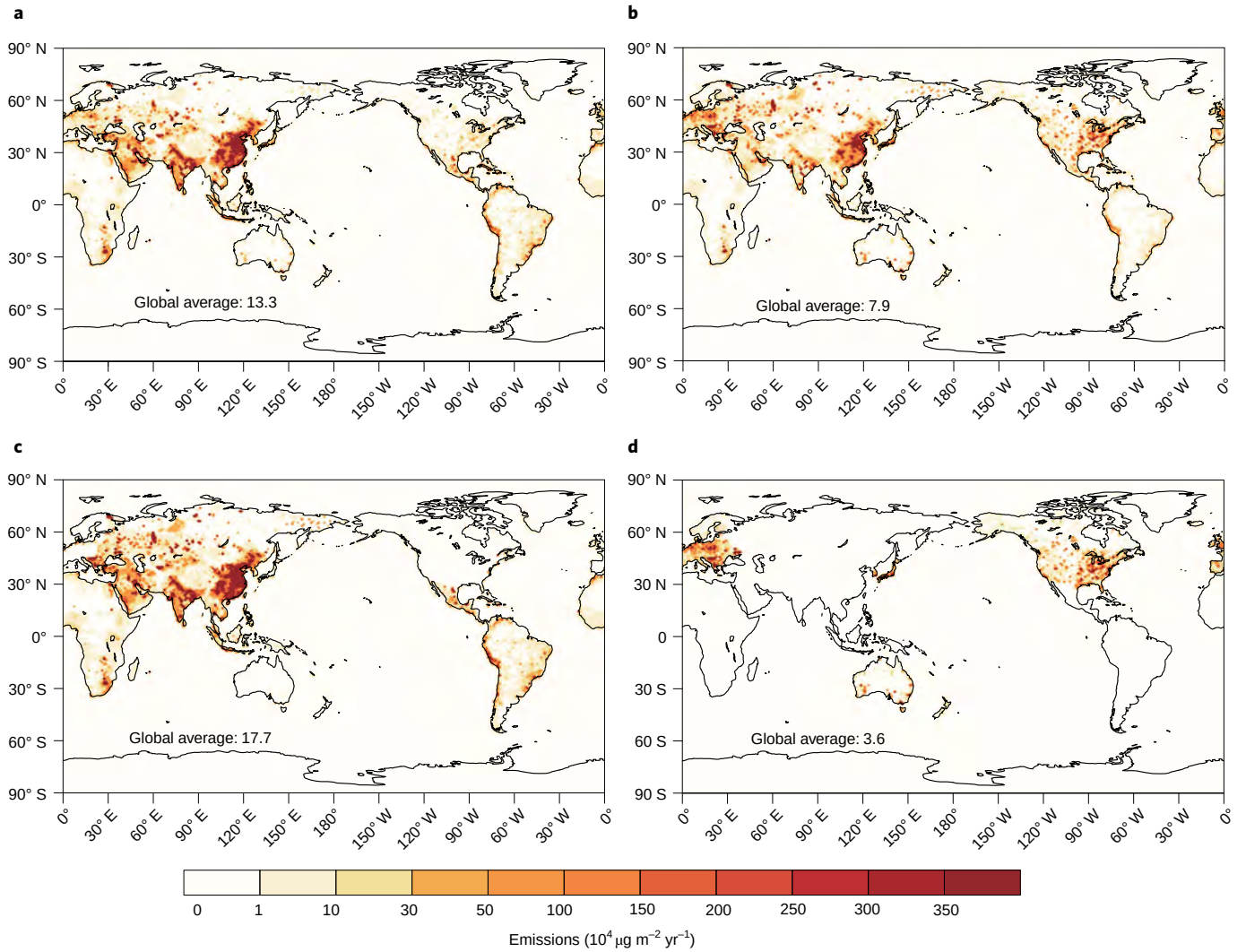



**Fig. 1 | Contrasting consumption-based and production-based emissions.** Anthropogenic $SO_2$ emissions associated with consumption and production activities of developing and developed countries in 2014. **a**, $E_c$ of developing countries. **b**, $E_c$ of developed countries. **c**, $E_p$ of developing countries. **d**, $E_p$ of developed countries. The $E_c$ data are calculated in this study, and the $E_p$ data are taken directly from the inventory. The global average is shown in each panel. Map outlines were generated using NCL scripts.

developed countries is based on International Monetary Fund[21]. Uncertainty estimates for global average results are represented by 2 s.d. (corresponding to 95% confidence interval (CI)), and a paired $z$ test for statistical significance is used to account for autocorrelation[22]. Our results show unexpectedly similar effects of consumption by developed and developing countries on the global average near-surface air temperature and precipitation, despite the large difference in the magnitude of consumption-associated $SO_2$ emissions by a factor of 1.7.

## Consumption-associated sulfur emissions and loadings

Figure 1 compares the production-based emissions ($E_p$) and consumption-based emissions ($E_c$) of $SO_2$ for both developed and developing countries. The $E_c$ of a region denote worldwide emissions due to production of goods and services to supply consumption by that region. By contrast, the $E_p$ are associated with production by that region to supply global consumption, as recorded in the inventory used here. The global total $E_c$ of developing countries (67.7 Tg $yr^{-1}$) are about 1.7 times those of developed countries (39.9 Tg $yr^{-1}$). This is in contrast to the factor-of-five difference in global total $E_p$ (89.7 versus 17.8 Tg $yr^{-1}$). The $E_c$ exhibit much greater spatial coverage than do the $E_p$, especially for the emissions of developed countries (Fig. 1).

Figure 1 further shows that the $E_c$ of developed countries that are released in developing countries are larger than the $E_c$ of developing countries that are released in developed countries by a factor of 7. These differences are because developed countries are net importers of goods from developing countries and have much lower emission intensities than the latter[6,23]. Overall, the $E_c$ of developing countries are more concentrated in Asia low and middle latitudes, whereas the $E_c$ of developed countries more spread out over northern latitudes. These geographical characteristics are further depicted in the zonal average and meridional average emissions in Extended Data Fig. 1. The unique features of regional $E_c$ in magnitude and spatial pattern warrant a detailed assessment of resulting climate impacts to supplement prevailing production-based studies.

Extended Data Fig. 2a,b shows the spatial distribution of sulfate aerosol column densities associated with consumption ($S_c$) by developed and developing countries. Sulfate is formed in the atmosphere from emitted $SO_2$ and lost through dry and wet deposition. The spatial pattern of $S_c$ largely follows the $E_c$ due to short atmospheric lifetimes of $SO_2$ and sulfate, although portions of sulfate

**Table 1 | Climate response to $E_c$ of $SO_2$**

| | Due to $E_c$ of developing countries | Due to $E_c$ of developed countries |
|---|---|---|
| $E_c$ ($SO_2$ $10^4$ μg m$^{-2}$ yr$^{-1}$) | 13.3 | 7.9 |
| $S_c$ ($SO_4$ μg m$^{-2}$) | 17.5 (±0.83) | 9.1 (±0.89) |
| ΔAOD ($10^{-2}$) | 0.50 (±0.32) | 0.32 (±0.36) |
| ERF (W m$^{-2}$) | -0.48 (±0.64) | -0.42 (±0.64) |
| Δ$T$ (K) | -0.18 (±0.11) | -0.20 (±0.09) |
| Δ$P$ (mm d$^{-1}$) | -0.019 (±0.019) | -0.017 (±0.017) |

Included are the changes in global average annual mean $S_c$, AOD, ERF, 2 m air temperature ($T$) and precipitation ($P$). The values in brackets represent 2 s.d.

are transported from emission source regions to downwind areas[3]. The global average $S_c$ of developing countries are 1.9 times the $S_c$ of developed countries, consistent with the difference in $E_c$. The $S_c$ of developing and developed countries are both large in the Northern Hemisphere (NH), but their regional patterns differ. The $S_c$ of developing countries are concentrated over East and South Asia. By comparison, the $S_c$ of developed countries are more evenly distributed in the zonal direction at northern latitudes, with high values over Eurasia and North America. The difference in spatial pattern is also evident for aerosol optical depth (AOD, Extended Data Fig. 2c,d).

## Climate responses to sulfur emissions

The $E_c$ of developing and developed countries reduce the global average annual mean 2 m air temperature by 0.18 ± 0.11 K and 0.20 ± 0.09 K, respectively (Table 1). The magnitudes of temperature reduction are both close to the magnitude of observed surface warming (~0.2 K) over the past decade[2]. The comparable temperature reductions are despite the global average $E_c$ and $S_c$ of developing countries being 1.7 and 1.9 times those of developed countries. The temperature reduction is most evident in the NH (Fig. 2a,b), by a hemispheric average of 0.23 K driven by the $E_c$ of developing countries and by 0.25 K driven by the $E_c$ of developed countries. The $E_c$ of developed countries cause significant cooling at the NH extratropics (30°–90° N) with weak effects over tropical East and South Asia. By comparison, the $E_c$ of developing countries cool East and South Asia but less so at higher latitudes of Eurasia (Fig. 2a). The zonal average results in Fig. 3a,b further show substantial cooling between 60° S and 60° N by the $E_c$ of each region (by 0.10–0.40 K depending on the latitude). The temperature responses in the polar areas are much more uncertain (Fig. 3) due to natural variability[14,22], although the Arctic cooling induced by the $E_c$ of developed countries are statistically significant at 68% CI.

The unexpectedly comparable cooling effects by the $E_c$ of developing and developed countries are because the $E_c$ of developed countries, although smaller in magnitude, have a stronger per-emission capability to induce forcing. To demonstrate this point, we calculate the effective radiative forcing (ERF) of $E_c$ with additional Community Atmosphere Model version 6 (CAM6) simulations (Methods). The global mean ERFs due to the $E_c$ of developing and developed countries are comparable (−0.48 ± 0.64 versus −0.42 ± 0.64 W m$^{-2}$), consistent with the similar temperature changes. The ERFs per emission of $SO_2$ for these two regions are −0.0071 ± 0.0094 and −0.011 ± 0.016 W m$^{-2}$ (Tg yr$^{-1}$)$^{-1}$, respectively, with the latter higher by about 50%.

Extended Data Fig. 3a,b further shows the horizontal distribution of ERF. The $E_c$ of developed countries lead to significant negative ERF throughout the NH middle and high latitudes, whereas the $E_c$ of developing countries result in the greatest ERF at the low and middle latitudes around China and India. The negative ERF caused by $E_c$ of developing countries is statistically significant at 95% CI over South Asia, East Asia and the downwind northern Pacific, significant at 68% CI over the majority of the NH but insignificant over the contiguous United States (Extended Data Fig. 3a). The negative ERF by $E_c$ of developed countries is statistically significant at 95% CI over parts of East Asia, the northern Pacific, northern Eurasia and North America and significant at 68% CI over many other NH areas (Extended Data Fig. 3b).

Decomposition of the ERF by using radiative kernels[24] shows major contributions from the direct radiative forcing of sulfate and fast adjustment of clouds, with minor effects from fast adjustments of other climate variables (Extended Data Table 1). The spatial distribution of sulfate direct radiative forcing (Extended Data Fig. 3c,d) follows those of $S_c$ and AOD (Extended Data Fig. 2), with the global average sulfate forcing induced by the $E_c$ of developing countries 2.5 times that of developed countries (−0.16 ± 0.031 versus −0.066 ± 0.026 W m$^{-2}$). The $E_c$ of developing and developed countries exert similar global average radiative effects via fast cloud adjustment (−0.32 ± 0.70 versus −0.32 ± 0.72 W m$^{-2}$, Extended Data Table 1). The fast cloud adjustment corresponds to the overall increase in clouds coverage (statistically significant at 95% or 68% CI over large regions; Extended Data Fig. 4).

Extended Data Table 2 shows radiative kernel decomposition[24] for the top-of-atmosphere (TOA) net radiative flux resulting from $E_c$ perturbation, as averaged over the past 110 years of fully coupled CESM2 simulations. Contributors to the net flux include ERF and the resulting responses of water vapour, surface albedo, temperature and clouds. The contributions of all non-ERF factors are in addition to the radiative effects of their fast adjustments presented in Extended Data Table 1. The 110-year average net fluxes due to the $E_c$ of developing and developed countries are much smaller than the respective values of ERF. Over the 110 years, the net flux is varying but with no trends (Supplementary Fig. 3). The net flux is not zero because the response of deep ocean to ERF (which may take thousands of years) has not been fully realized; this is evident from the imbalance in surface energy flux over the oceans (Extended Data Fig. 3e,f).

Extended Data Table 2 shows that the ERF-induced reduction in surface temperature (in addition to the small fast adjustment) leads to a large positive radiative effect, whereas the corresponding reduction in water vapour leads to a negative radiative effect. The ERF-induced change in clouds (in addition to the fast adjustment) leads to a small radiative effect. The total radiative effect of clouds change induced by $E_c$ of each region is negative (−0.23 ± 1.26 W m$^{-2}$ for developing countries and −0.26 ± 1.43 W m$^{-2}$ for developed countries), calculated as the sum of the respective values in Extended Data Tables 1 and 2. This is consistent with the overall enhancement in clouds coverage simulated by fully coupled CESM2 (statistically significant at 95% or 68% CI over vast areas; Extended Data Fig. 5).

In the Arctic areas, the $E_c$ of developed countries lead to an increase in sea-ice coverage (Extended Data Fig. 6b) and a deepened polar vortex (as manifested in the geopotential height and wind field at 500 hPa, Extended Data Fig. 7b). The increase in sea ice coincides with the strong surface cooling there (Fig. 2b). The change in polar vortex is consistent with the widespread cooling over the northern middle and high latitudes (Fig. 2b). By comparison, the $E_c$ of developing countries lead to weaker Arctic responses (Extended Data Figs. 6a and 7a) because the emissions are located at lower latitudes.

The $E_c$ of developing and developed countries each result in warming along the coasts of Greenland. The temperature change is consistent with the reduction in sea-ice coverage (Extended Data Fig. 6a,b) and the increase in the surface (latent + sensible) heat flux to the atmosphere (Extended Data Fig. 6e,f). Correspondingly, the change in surface winds reduces the sea-ice drift to these locations (Extended Data Fig. 6c,d).

In the tropics, the Southern Hemisphere (SH) Hadley cell circulation strengthens and the NH Hadley cell weakens, and the associ-

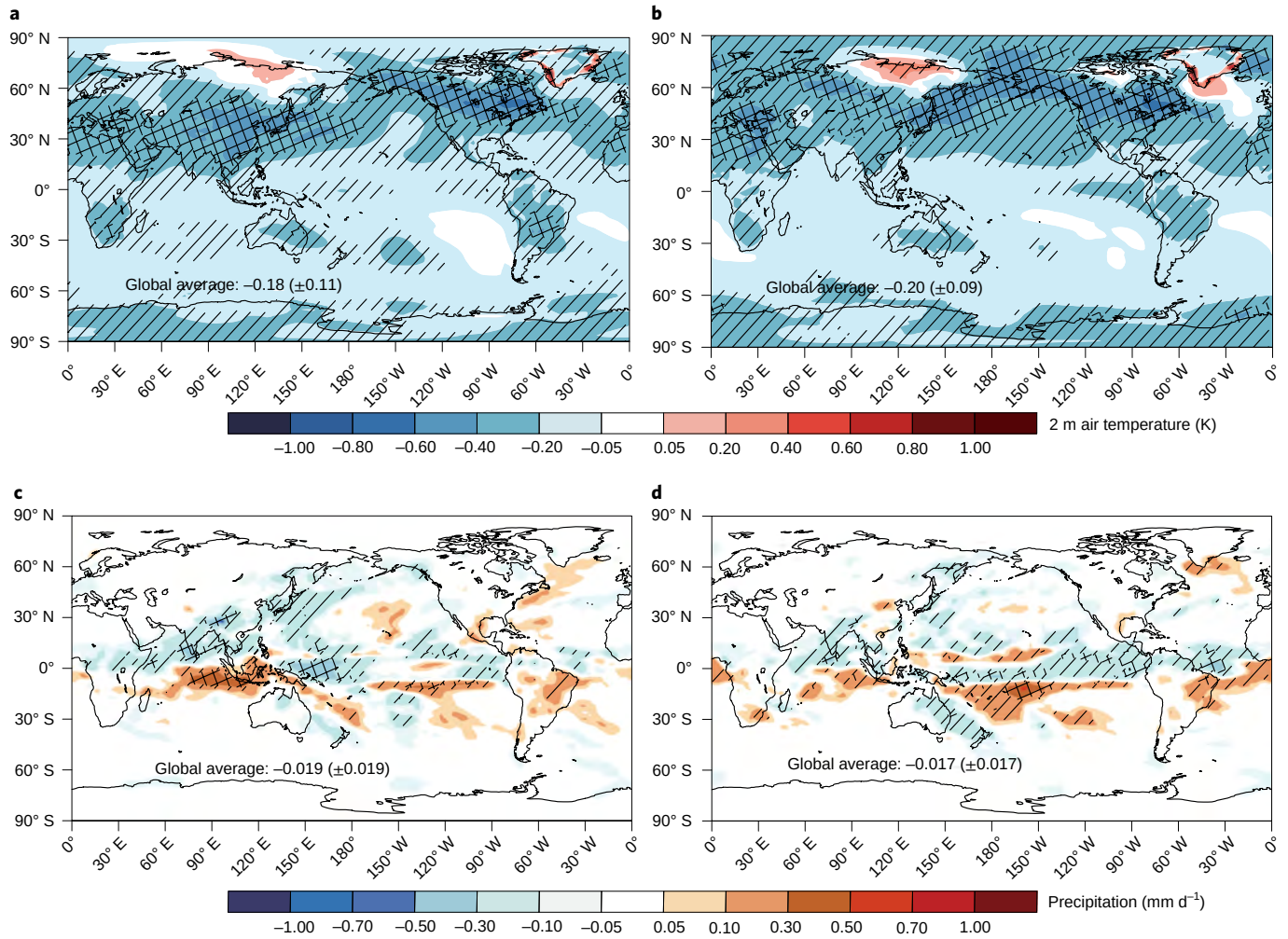



**Fig. 2 | Climate response to consumption-based emissions. a**, Changes in annual mean 2 m air temperature due to $E_c$ of developing countries. **b**, Changes in annual mean 2 m air temperature due to $E_c$ of developed countries. **c**, Changes in precipitation due to $E_c$ of developing countries. **d**, Changes in precipitation due to $E_c$ of developed countries. The grid cells with cross lines pass the paired *z* test at 95% CI, and those with diagonal lines pass the test at 68% CI. The global average is shown in each panel. Map outlines were generated using NCL scripts.

ated intertropical convergence zone shifts southwards, as manifested in the weakened rising in the NH tropics and enhanced rising in the SH tropics (Extended Data Fig. 8). These tropical responses are because the sulfate-induced cooling is stronger in the NH than in the SH, resulting in reduced hemispheric temperature asymmetry (because the NH is warmer than the SH climatologically) and less atmospheric cross-Equator energy transport to the SH by the Hadley cell[25,26]. The tropical circulation change is associated with weakened precipitation[27,28] in South and East Asian monsoon areas (Fig. 2c,d), which has a warming effect[29] that partly offsets the radiative cooling effect. Since the magnitude of ERF by developing countries is large in East and South Asia (Extended Data Fig. 3a), the sinking enhancement at 0–20° N is strong (Extended Data Fig. 8a). The ERF by developed countries occurs at higher latitudes and is zonally evener. The resulting effect on the Hadley cell is weaker in magnitude (Extended Data Fig. 8b versus 8a), as also reflected in the weaker change in meridional winds in tropical Asian monsoon areas at 850 hPa (Extended Data Fig. 7d versus 7c).

The $E_c$ of developing and developed countries result in similar reductions in global average annual mean precipitation ($-0.019 \pm 0.019$ versus $-0.017 \pm 0.017$ mm d$^{-1}$, Fig. 2c,d) since the response of precipitation is strongly constrained by radiative energy balance[30]. The calculated changes in precipitation are statistically significant at 68% CI over the majority of tropical oceans. The precipitation changes are even statistically significant at 95% over parts of the tropical regions of China, India, the Indian Ocean and the western Pacific due to the $E_c$ of developing countries (Fig. 2c) and over parts of the tropical Pacific due to the $E_c$ of developed countries (Fig. 2d). The $E_c$ of developing countries result in the greatest reduction (by up to 0.57 mm d$^{-1}$, statistically significant at 95% CI; Fig. 2c) over tropical Asian continental monsoon areas (India, IndoChina Peninsula and south China), whereas the $E_c$ of developed countries do not cause such a strong reduction (Fig. 2c,d). The zonal average results also show the strongest precipitation changes in the tropics (by up to about $\pm 0.1$ mm d$^{-1}$), with a reduction in the NH and an enhancement in the SH (Fig. 3c,d), consistent with the changes in the Hadley cell and intertropical convergence zone (Extended Data Fig. 8). The peak zonal average precipitation enhancement over the SH tropics is statistically significant at 68% CI.

## A consumption-based perspective for climate action

For each region, there exist substantial differences between its $E_c$ and $E_p$ in the magnitude and geographical pattern. The climate response to emissions associated with regional consumption presented in this study is very different from that inferred from prevailing production-based studies. For example, applying the global

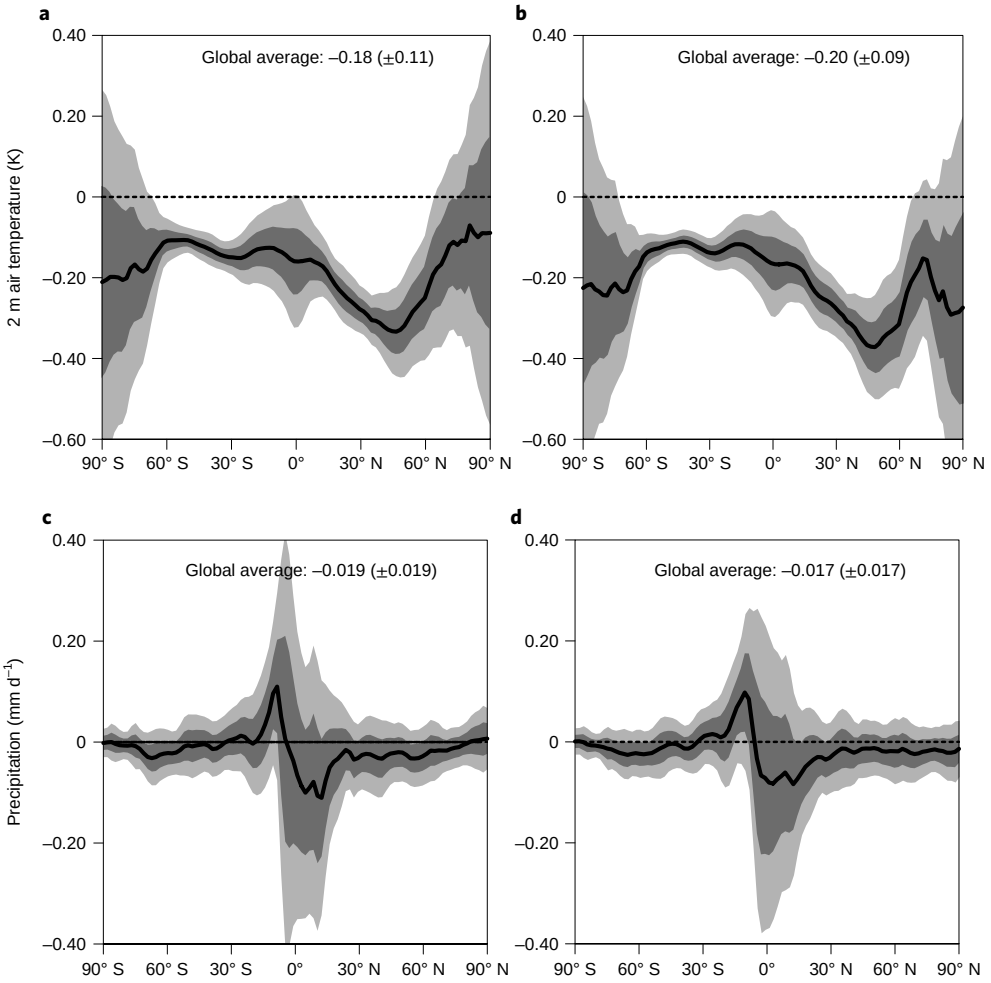



**Fig. 3 | Zonal average climate response to consumption-based emissions. a**, Changes in zonal average annual mean 2 m air temperature in response to $E_c$ of developing countries. **b**, Changes in zonal average annual mean 2 m air temperature in response to $E_c$ of developed countries. **c**, Changes in precipitation in response to $E_c$ of developing countries. **d**, Changes in precipitation in response to $E_c$ of developed countries. The light shading indicates 2 s.d. of the 110 yr time series, and the dark shading indicates 1 s.d. The global average value is shown in each panel.

average temperature change per unit of $E_p$ suggested by ref. [9] would mean a factor-of-five difference in the temperature response to the $E_p$ of developing and developed countries, in contrast to the comparable temperature effects by their $E_c$ shown here. This suggests that a picture of the regional contribution to climate change drawn from both production and consumption perspectives will be more complete in supporting regional efforts to combat climate change.

Although the $E_c$ of $SO_2$ by developing and developed countries differ by a factor of 1.7, they result in comparable changes in the global average surface air temperature and precipitation. This highlights the complex response of the nonlinear climate system to not only the magnitude but also the spatial pattern of aerosols. This finding is broadly consistent with previous $E_p$-based, idealized model simulations[8–13], although our results are based on realistic $E_c$ emissions. For climate policy considerations and international negotiations, a climate change attribution assessment based on fully coupled model simulations driven by realistic, regionally detailed emissions is important. Such an assessment will also offer a reference to validate and correct results based on simplified metrics (for example, global average radiative forcing and global warming potential) in situations where these metrics are used to save computational resources and time.

Our study is based on emissions in a recent year. Along with the economic globalization and trade liberation, the world has experienced rapid changes in regional $E_p$ and $E_c$. The $E_p$ of developed countries have continued to decline for decades[18], but their $E_c$ have declined at much weaker rates[3,31]. Over the recent few years, the $E_p$ of China have been reduced substantially[32], whereas the $E_p$ of many developing countries have continued to grow or changed little[33,34] (except for the temporary disruption by COVID-19), and the resulting changes in $E_c$ remain unclear. Future changes in $E_p$ and $E_c$ will depend on the pathway of socioeconomic development and emission control. Many future scenarios assume transition in the energy–industrial sector at a scale not experienced in the past[2,35]. A few ongoing events such as the Sino-US trade war[23], Brexit and COVID-19 have disrupted the global trade, but the duration of these events and their direct and indirect emissions impacts are less clear. Also not clear are whether the world will embrace continuous globalization—regional rivalry is one of the few plausible shared socioeconomic pathways projected for this century[36]—and how this will affect the magnitude and distribution of regional $E_c$. These pose important questions of how the continuously changing $E_c$ of each region have been affecting and will affect the climate system.

Our study is focused on the effect of sulfate. However, consumption activities also affect other aerosols (for example, black carbon) and greenhouse gases. Given the importance of responsible consumption to the success of climate change mitigation and environmental protection[1,2], it will be very beneficial to conduct a com-

prehensive assessment of the climate response to the whole suite of climate-relevant constituents associated with regional consumption. A multimodel assessment will help reduce the intrinsic uncertainty in model simulations. This, of which our sulfate study serves as the first step, will complement prevailing production-based analyses to help understand how human activities affect the climate system more thoroughly. As much of the $E_c$ of any region are released in other regions, such a consumption-based assessment will also provide a basis for international cooperation to act towards sustainable consumption and climate change mitigation, which will have substantial co-benefits for public health[4].

## Online content

Any methods, additional references, Nature Research reporting summaries, source data, extended data, supplementary information, acknowledgements, peer review information; details of author contributions and competing interests; and statements of data and code availability are available at https://doi.org/10.1038/s41561-022-00898-2.

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## Methods

**An interdisciplinary framework.** Through an interdisciplinary framework, we employ sector-, region- and pollutant-specific anthropogenic emission data from the Community Emissions Data System (CEDS) inventory for the Coupled Model Intercomparison Project phase 6 (CMIP6) project[18], use the latest Global Trade Analysis Project (GTAP) multiregional input–output (MRIO) table (version 10 for 2014)[19] to estimate consumption-related emissions of $SO_2$, and run the fully coupled CESM2[20] to simulate the climate impacts of such sulfur emissions. The widely used[4,6,23] GTAP MRIO table offers quantitative information on the inter-sectoral and inter-regional economic linkage along the global supply chain and tracks worldwide production activities in support of consumption of each region. The MRIO table together with the CEDS CMIP6 emission inventory allows for an estimate of consumption-based emissions. Such emissions are then implemented into the fully coupled CESM2 for climate impact simulations, supplemented by additional atmosphere-only simulations.

**The CESM2 Model.** We use fully coupled CESM2 in this study. CESM2 is the most current-generation Earth system model developed by the National Center for Atmospheric Research and other institutions. CESM2 is built on many successes of its predecessor, CESM1.2, which is one of the best models in the CMIP5 set of model simulations according to a pattern distance metric of present-day precipitation and temperature[37,38]. CESM1.2 captures the spatial pattern of climatic surface temperature at a country scale in the meteorological reanalysis data, with a high correlation coefficient and a low normalized mean deviation[39]. CESM2 contains many updates to CEMS1.2 and results in an improved historical simulation, including a better representation of low-latitude precipitation and shortwave cloud forcing[20]. An overview of CESM2 can be found elsewhere[20].

CAM6 is the atmospheric component of CESM2. Inside CAM6, we choose a fully interactive aerosol scheme, the Modal Aerosol Model version 4 (MAM4)[40], following the CESM2 set-up for submission to CMIP6. The main difference between MAM4 and the MAM3 scheme used in CESM1.2 is the representation of carbonaceous aerosols[40]. $SO_2$ is oxidized by the hydroxyl radical, hydrogen peroxide and ozone to form sulfate ($SO_4$), and sulfate size distributions are simulated prognostically in three lognormal modes: Aitken, accumulation and coarse[41,42]. Both MAM4 and MAM3 provide aerosol simulations with reasonable agreement with observations[43,44]; and they have been used widely to simulate the effects of sulfate on global temperature, precipitation and local climate[11,14,22,45–51]. In this study, CAM6 is run at a horizontal resolution of 1.9°×2.5° with 32 vertical levels.

CESM2 includes an interactive land surface model (the Community Land Model version 5)[52] run at the same horizontal resolution as CAM6, with an extensive suite of new and updated processes and parameterizations to previous versions. The ocean component of CESM2 adopts the Parallel Ocean Program version 2 (POP2)[53] with several physics and numerical improvements[20] and is run at an equivalent resolution of 1°. CESM2 uses the Los Alamos sea-ice model version 5.1.2 (CICE5)[54] as its sea-ice component, and CICE5 shares the same horizontal grid with POP2.

Supplementary Fig. 1a describes the suite of model simulations in this study. We conduct three cases of fully coupled CESM2 simulations, including GLOBAL, Xdeveloping_CE and Xdeveloped_CE. The three cases differ only in anthropogenic $SO_2$ emissions, with global emissions in GLOBAL, removal of $E_c$ of developing countries in Xdeveloping_CE and removal of $E_c$ of developed countries in Xdeveloped_CE. The impacts of consumption by developing and developed countries are obtained by subtracting simulation results of Xdeveloping_CE and Xdeveloped_CE from GLOBAL, respectively.

Before running the three cases, we conduct an additional fully coupled simulation Pre_run for 300 years from an initial climate state in the year 2000. The GLOBAL case then starts from the climate state in the 300th year of the Pre_run and continues for an additional 407 years; Supplementary Fig. 2 shows that the climate has reached quasi-equilibrium state at this time. Xdeveloping_CE contains three ensemble members branched from the 201st, 211th and 265th years of GLOBAL, respectively. Xdeveloped_CE contains three ensemble members branched from the 201st, 235th and 247th years of GLOBAL, respectively. Each ensemble member of Xdeveloping_CE and X_developed_CE is run for 140 years, with the last 110 years used for analysis. Depending on the years of each ensemble member of Xdeveloping_CE and X_developed_CE, results from GLOBAL are paired temporally for comparison.

Each simulation outputs monthly mean values for temperature, precipitation and other variables. For Xdeveloping_CE or X_developed_CE, each of the three ensemble members provides a monthly time series of paired difference relative to GLOBAL, and the three resulting time series are averaged to obtain an ensemble mean time series. The 110 yr average and standard deviation for the ensemble mean time series are analysed. To correct for the model autocorrelation, we apply the paired $z$ test[22] to the monthly time series and focus the analysis on areas where the climate response passes the significance test with respect to 95% or 68% CI.

**Effective radiative forcing calculations.** The ERF represents the change in the TOP net (shortwave + longwave) radiative flux due to emission perturbations and fast adjustments of the atmosphere and land surface, with sea surface temperature and sea ice unchanged.

To calculate the ERF, we conduct three additional simulations of CAM6 with specified sea surface temperature and sea ice, following ref. [22]. Here, the ERF is computed from individual simulations with $SO_2$ emissions identical to GLOBAL, Xdeveloping_CE or X_developed_CE. An all-emission CAM6 simulation is run for 199 years. Other simulations branch from the fifth year of the all-emission simulation and are run for 195 years. The last 190 years of all simulations are used to calculate the ERF.

**Radiative kernel decomposition.** We calculate the TOA radiative feedbacks using the radiative kernel method introduced by ref. [24], which has been widely used in decomposition of radiative effects for climate change analysis[55–61]. The radiative kernel, $K_X$, is pre-calculated by a partial perturbation method using the radiative transfer model RRTM[62] for temperature ($T$), water vapour ($q$) and surface albedo ($a$). Then, for non-cloud climate variables, the radiative feedback is computed as

$$\Delta_X R = K_X \Delta X$$

for $X = T$, $q$ or $a$. The direct radiative forcing ($G$) from sulfate is calculated on the basis of CAM6 simulations, as determined by contrasting the TOA radiative flux (calculated at each time step but outputted every month) from CAM6 simulations with and without the $E_c$ (GLOBAL versus Xdeveloping_CE or GLOBAL versus X_developed_CE). The cloud feedback is then obtained from the residual of the radiation budget

$$\Delta_{\mathrm{cld}} R = \Delta R - \mathrm{G} - K_T \Delta \mathrm{T} - K_q \Delta q - K_a \Delta a$$

where $\Delta R$ is the response of the net TOA radiative flux.

**Emissions.** In a globalized economy with regions and sectors inter-connected through trade, the difference between $E_p$ and $E_c$ lies in the attribution of emissions associated with production to supply consumption. For example, a cell phone sold in the United States may be assembled in China with steel produced in Japan and iron ore extracted in Australia, and all associated emissions are allocated to the United States from the consumption perspective ($E_c$) but to these individual countries from the production perspective ($E_p$).

Anthropogenic emissions of all species but $SO_2$ represent the year 2000, as taken from the CESM2 present-day simulations. Emissions of $SO_2$ for the year 2014 are obtained by combining the (production-based) CEDS CMIP6 inventory[18] and the latest version (v.10, for 2014) of the widely used[4,6,23] GTAP MRIO table[19].

The GTAP MRIO table is a result of an internationally collaborative effort to produce a regionally and sectorally harmonized, high-quality input–output database for global trade analysis. It has been widely used as a core dataset in general equilibrium models. The disaggregation details of regions and sectors in the GTAP v.10 table are compatible with those in CEDS, with sectoral and regional mapping established in previous work[23,63].

The calculation of $SO_2$ $E_c$ follows our previous studies[4,6,23] and is briefly described here (see the flow chart in Supplementary Fig. 1b). We first obtain the worldwide $E_p$ of $SO_2$ in 2014 from the CEDS CMIP6 emissions inventory, which specifies monthly emissions for 55 sectors on a 0.5°×0.5° horizontal grid. We also calculate the production and consumption (each in monetary quantity) in 57 sectors of 141 countries/regions on the basis of the GTAP v.10 MRIO table for 2014. Production of each country/region and sector is to supply direct and indirect consumption by all regions and sectors. GTAP provides quantitative information of the linkages between sectors and regions along the global supply chain. The mapping from 55 sectors in CEDS to 57 sectors in GTAP follows ref. [23] (see their supplementary data 6). We then calculate country- and sector-specific emission intensities (emission divided by monetary economic output) of $SO_2$ on the basis of the $E_p$ and production data. We further calculate the $E_c$ of individual countries/regions and sectors by multiplying the sectoral emission intensities with consumption-associated sectoral output. Within each country, the spatial and monthly distributions of $E_c$ follow the distributions of the corresponding $E_p$. For example, the $E_c$ of the United States that are released in China (due to Chinese production for US consumption) follow the spatial and monthly distributions of the $E_p$ of China. For consumption activities not associated with monetary economic output such as residential use and private vehicle driving (and thus not shown in the GTAP MRIO table), emissions are taken from the CEDS CMIP6 dataset directly, following ref. [23].

## Data availability

The GTAP data are available at https://www.gtap.agecon.purdue.edu/. The CEDS CMIP6 emissions are available at https://esgf-node.llnl.gov/search/input4mips/. Model results that support the findings of this study are available on Zenodo with the identifier https://doi.org/10.5281/zenodo.5722962. Raw model simulation results generated during this study are large in size and are available upon reasonable request. Source data are provided with this paper.

## Code availability

The CESM is available at http://www.cesm.ucar.edu/models/cesm2/. The codes for kernel calculation are available at http://www.meteo.mcgill.ca/~huang/. Data

analysis and visualizations are created by the NCAR Command Language, which is freely available at https://doi.org/10.5065/D6WD3XH5. All computer codes generated during this study are available from the corresponding authors upon reasonable request.

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## Author contributions

J.L. and G.H. conceived the study. J.L. and C.Z. designed the research. L.C. calculated the emissions. C.Z. conducted the CAM and CESM simulations. P.L., J.W. and Y.Y. contributed to the experiments. Y.X. conducted radiative kernel calculations. J.L., C.Z., G.H., J.-F.L., J.N., J.Y. and K.H. analysed the results. J.L. and C.Z. wrote the paper with comments from J.-F.L., J.N., J.Y., K.H. and Y.H.

## Competing interests

The authors declare no competing interests.

## Additional information

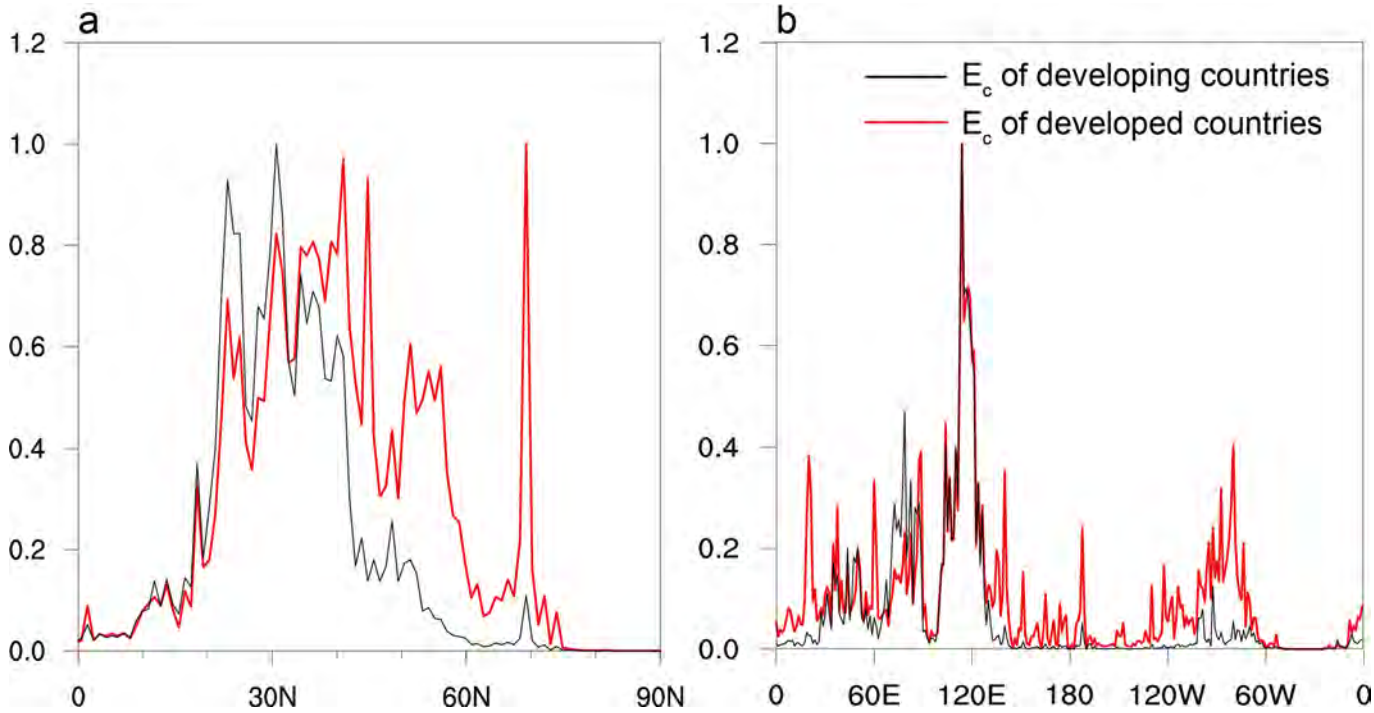
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**Supplementary information** The online version contains supplementary material available at https://doi.org/10.1038/s41561-022-00898-2.

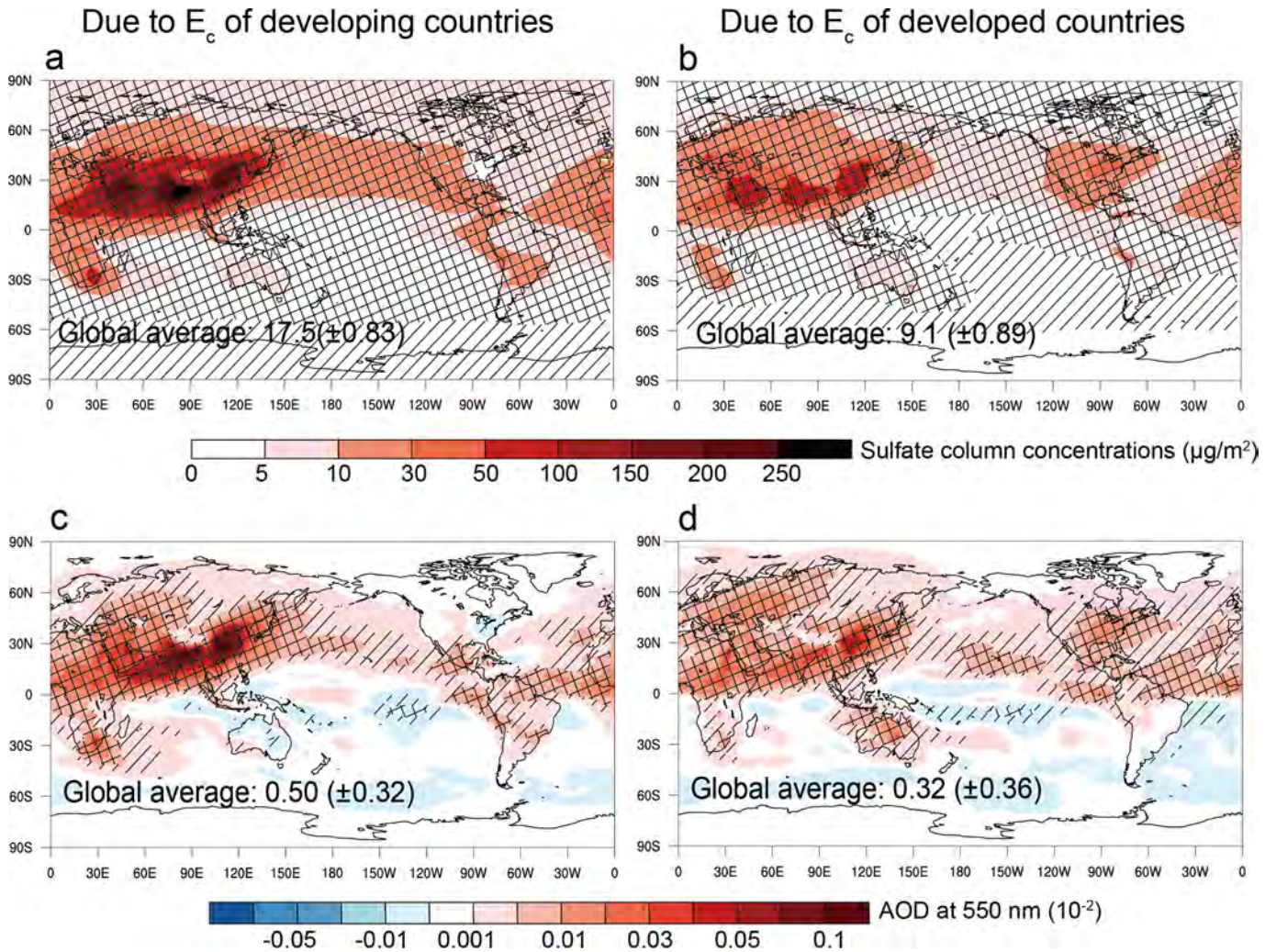
**Correspondence and requests for materials** should be addressed to Jintai Lin or Gang Huang.

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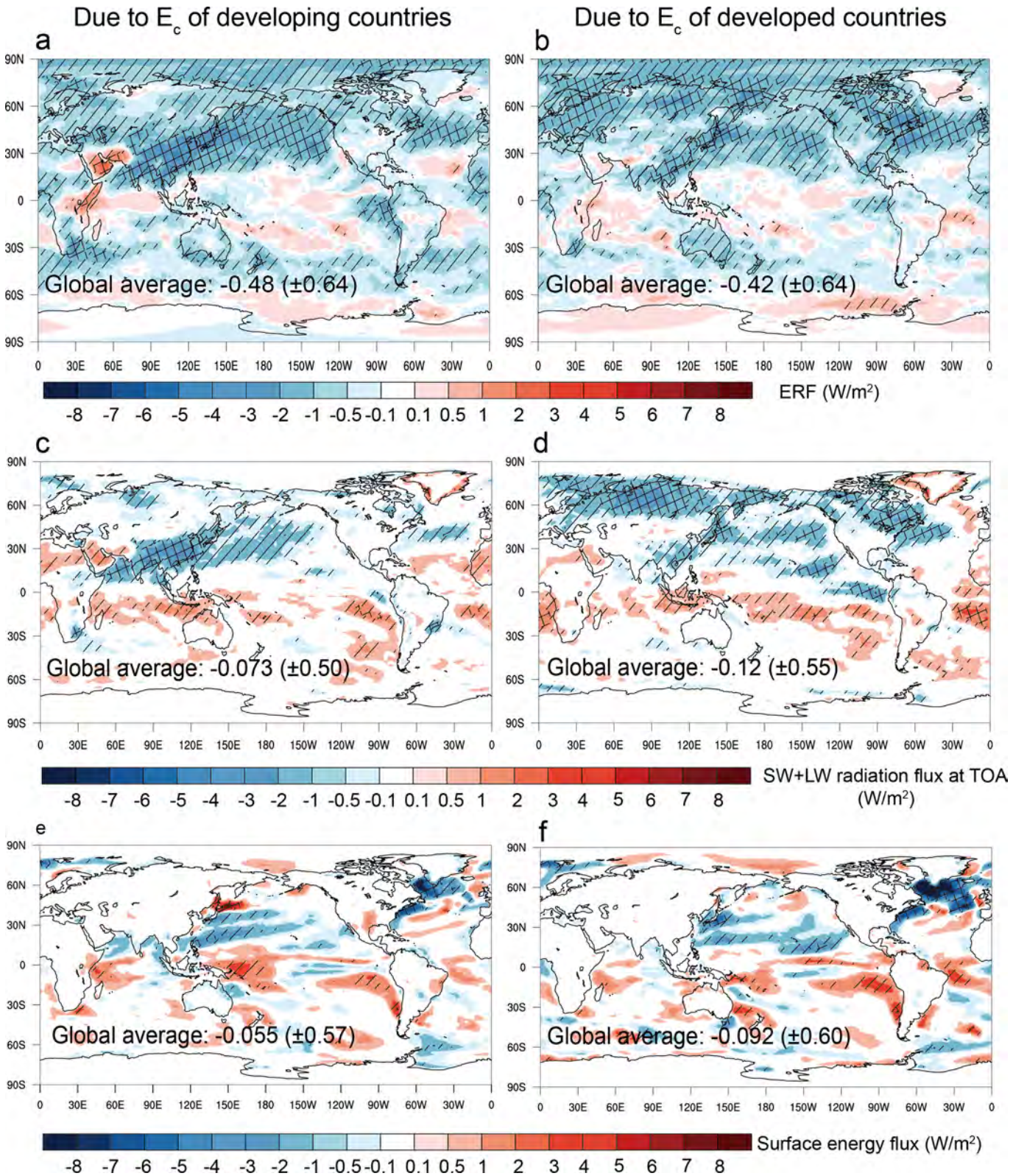
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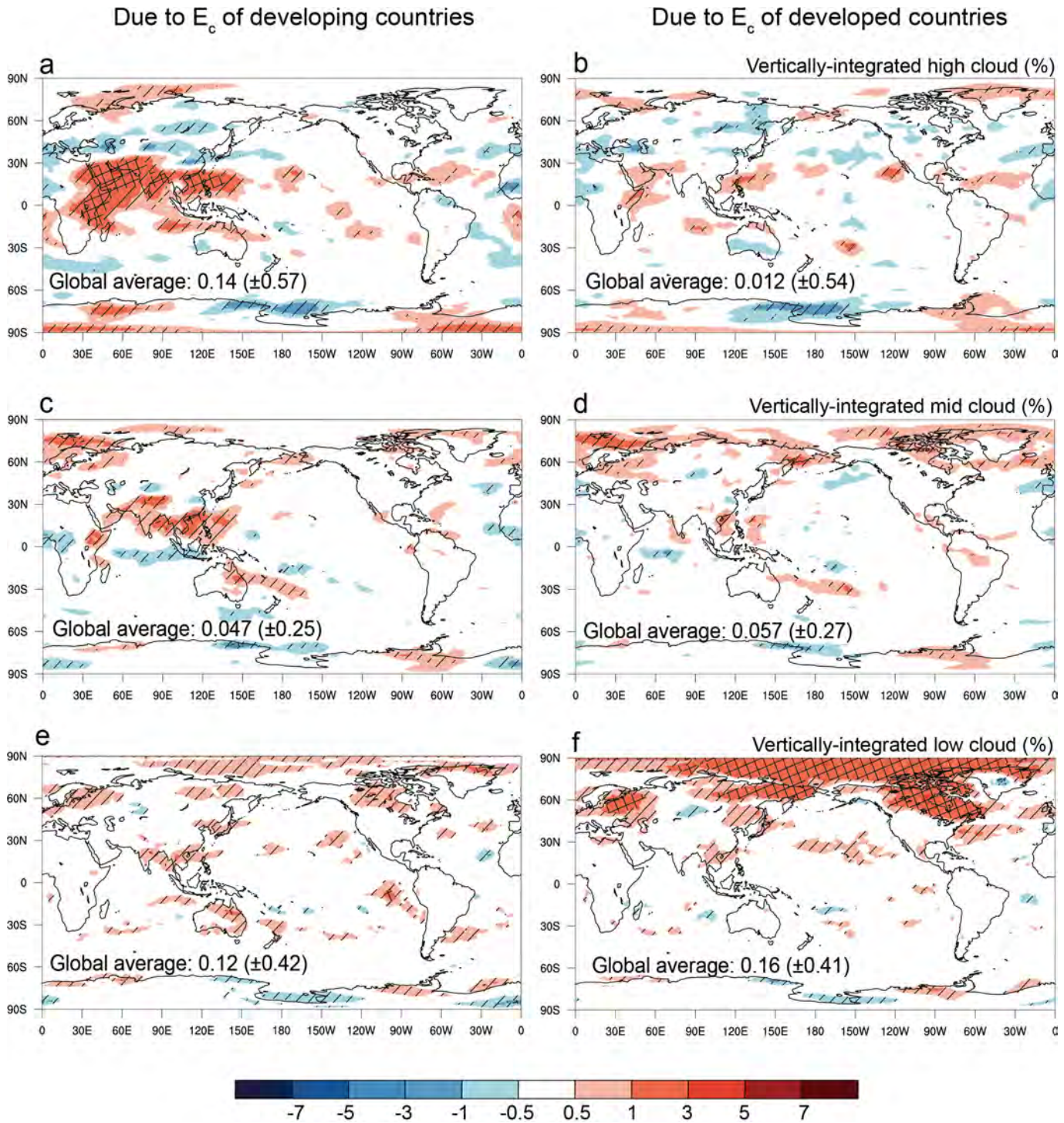

**Extended Data Fig. 1 | Latitudinal and longitudinal variations of emissions in the Northern Hemisphere.** Zonal average $E_c$ (a) by developing and developed countries, and meridional average $E_c$ (b). Emission values are averaged over land and then normalized relative to their min-max value.



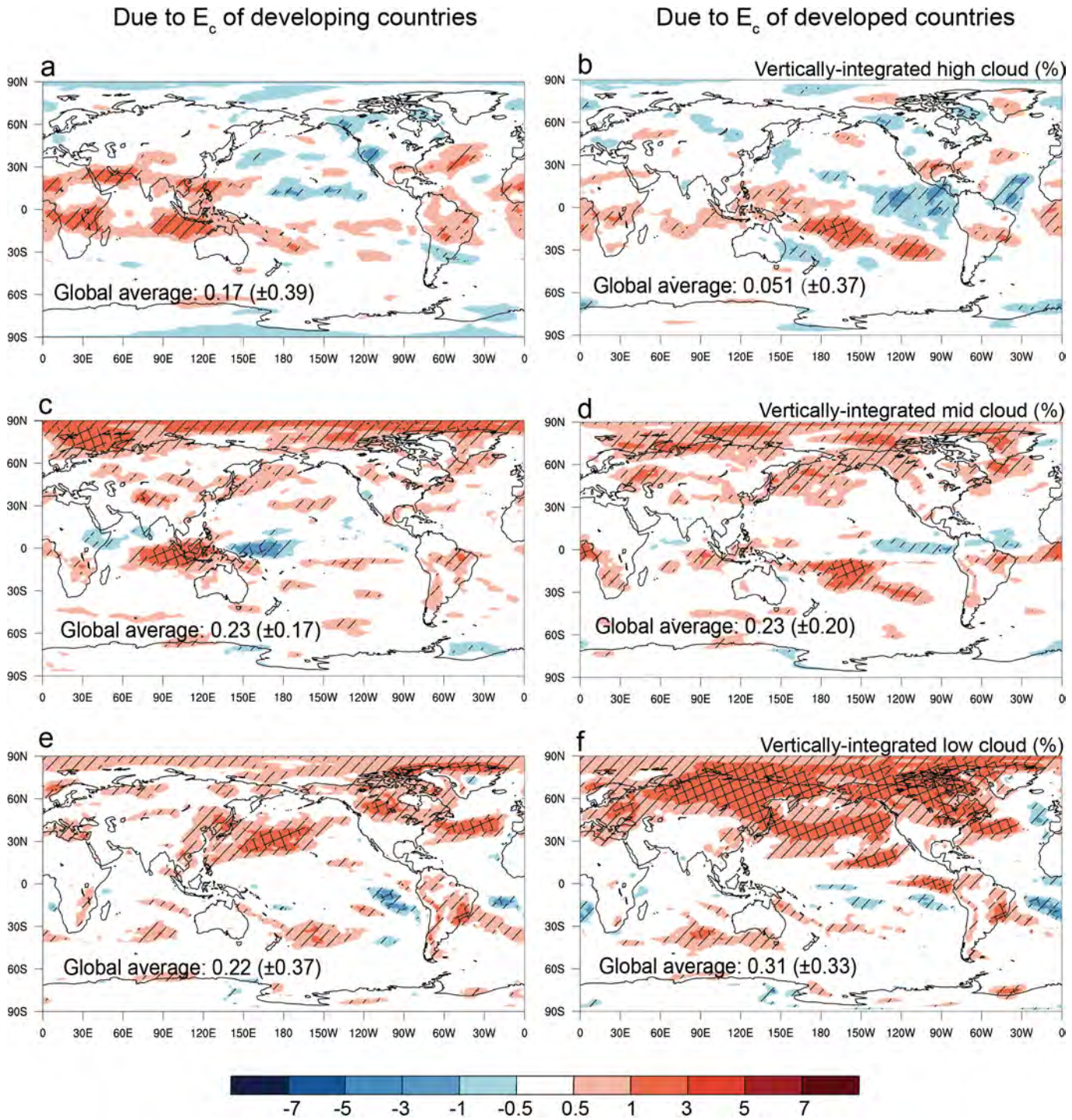

**Extended Data Fig. 2 | Sulfate and AOD responses to consumption-based emissions.** Changes in annual mean $S_c$ (a, b) and AOD (c, d) due to $E_c$ of developing and developed countries. AOD includes the effect of all aerosol compositions. The grid cells with cross lines pass the paired z-test at 95% CI and those with diagonal lines pass the test at 68% CI. The global average with 2 standard deviations is shown in each panel.



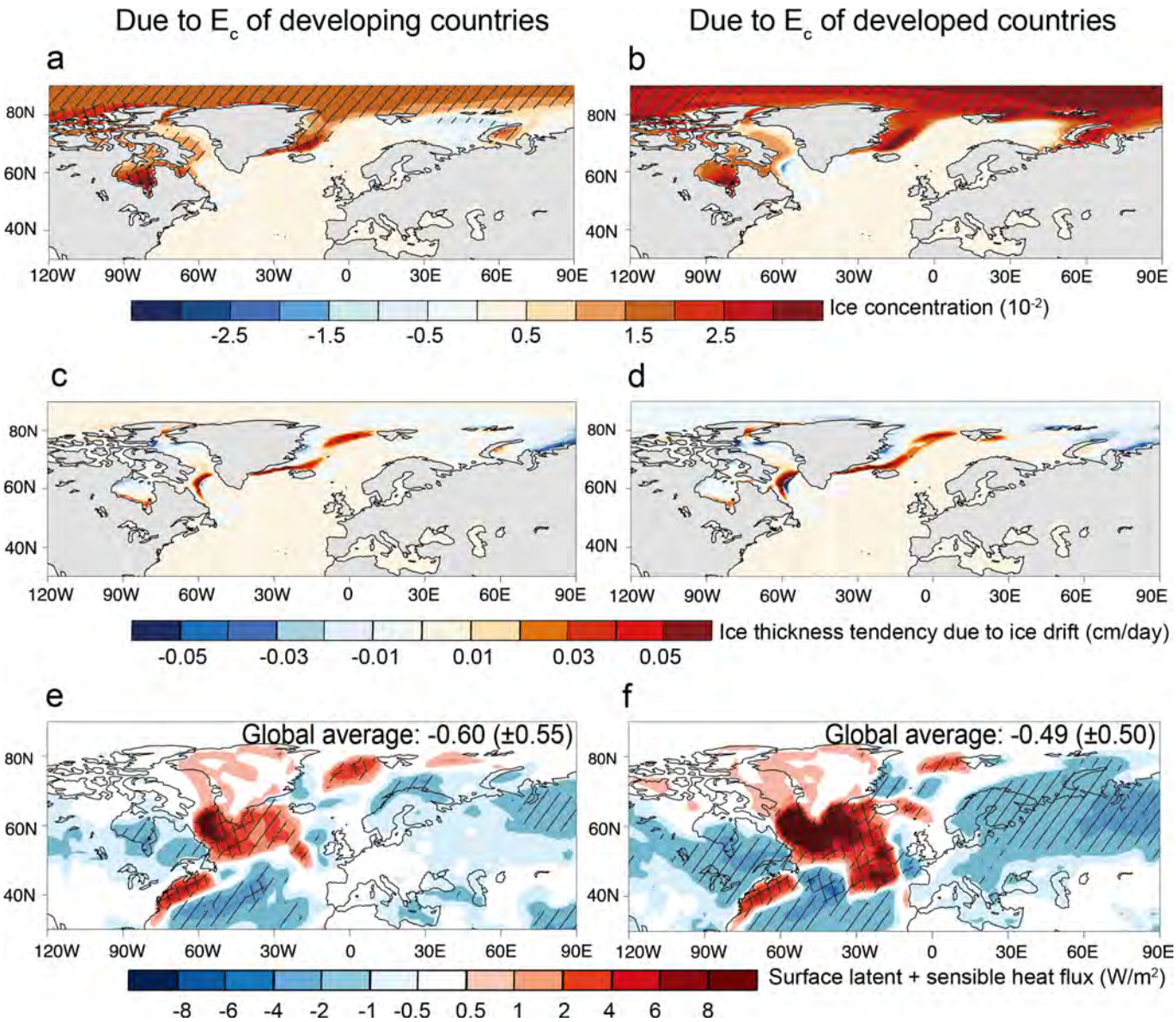

**Extended Data Fig. 3 | Energy imbalance induced by consumption-based emissions.** (a,b) Annual mean ERF due to $E_c$ of developing and developed countries calculated based on CAM6 simulations. (c-f) Annual mean TOA net radiative flux (shortwave + longwave; c,d) and surface energy flux (shortwave + longwave + latent heat + sensible heat; e,f) due to $E_c$ of developing and developed countries calculated based on fully coupled CESM2 simulations. The grid cells with cross lines pass the paired *z*-test at 95% CI and those with diagonal lines pass the test at 68% CI. The global average with 2 standard deviations is shown in each panel.



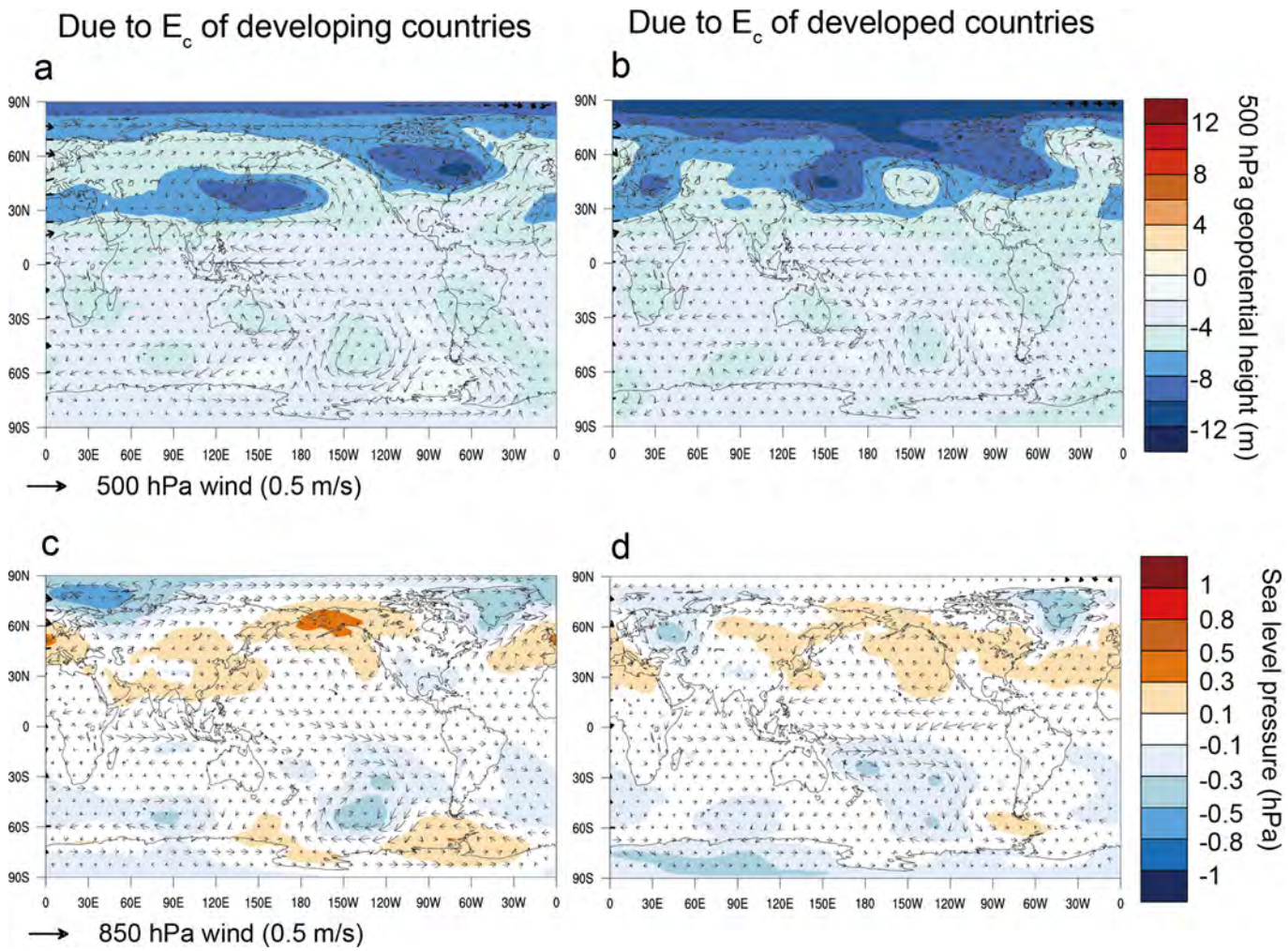

**Extended Data Fig. 4 | Fast adjustment of cloud coverage induced by consumption-based emissions in CAM6 simulations.** Changes in annual mean high clouds (cloud top pressure < 400 hPa), mid clouds (700–400 hPa) and low clouds (> 700 hPa) due to $E_c$ of developing and developed countries. The grid cells with cross lines pass the paired z-test at 95% CI and those with diagonal lines pass the test at 68% CI. The global average with 2 standard deviations is shown in each panel.



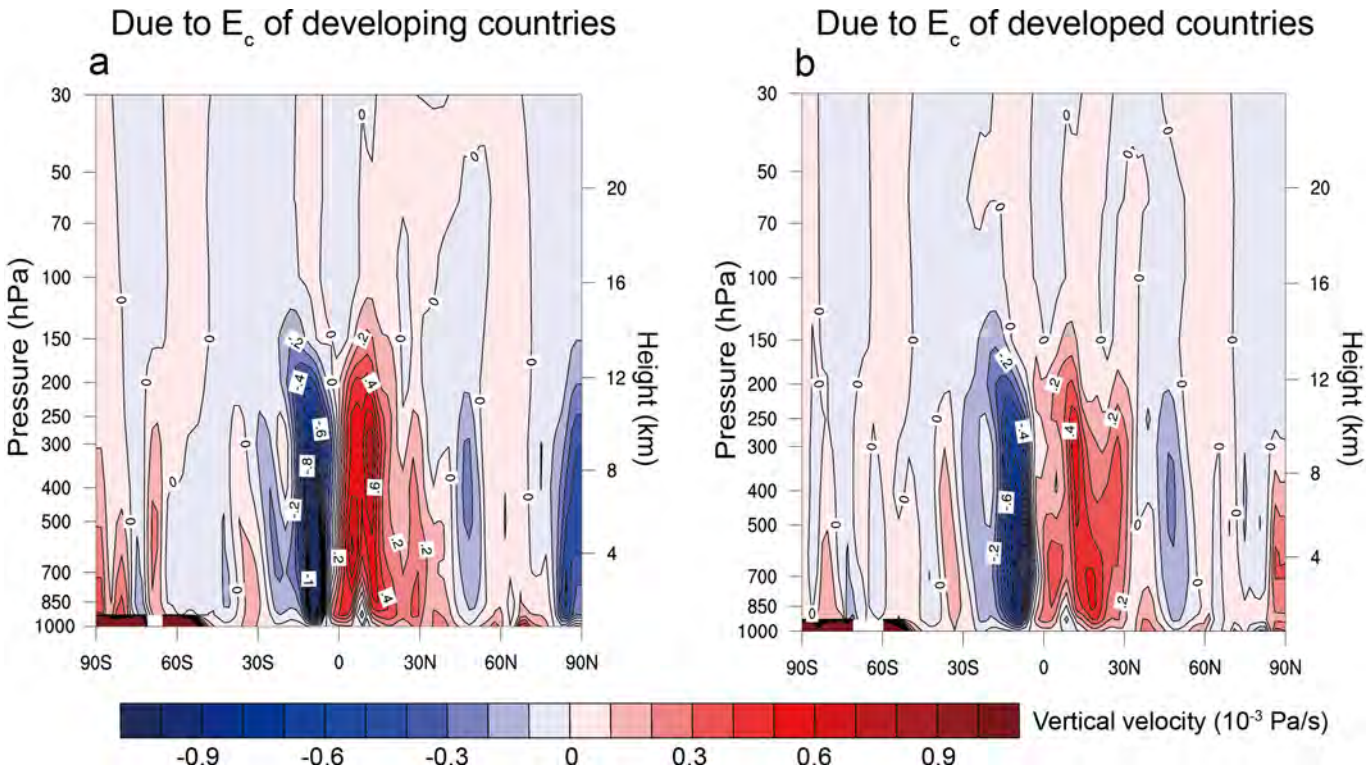

**Extended Data Fig. 5 | Response of cloud coverage to consumption-based emissions in fully coupled CESM2 simulations.** Changes in annual mean high clouds (cloud top pressure < 400 hPa), mid clouds (700–400 hPa) and low clouds (> 700 hPa) due to $E_c$ of developing and developed countries. The grid cells with cross lines pass the paired z-test at 95% CI and those with diagonal lines pass the test at 68% CI. The global average with 2 standard deviations is shown in each panel.




**Extended Data Fig. 6 | Responses of sea ice and heat flux to consumption-based emissions.** Changes in annual mean sea ice concentration (that is, fraction of the area covered by sea ice in each grid cell) (a, b), wind-driven sea ice drift (c, d) and surface heat flux (latent + sensible) (e, f) due to $E_c$ of developing and developed countries. The grid cells with cross lines pass the paired z-test at 95% CI and those with diagonal lines pass the test at 68% CI. In e and f, the global average of surface heat flux with 2 standard deviations is shown for comparison.




**Extended Data Fig. 7 | Circulation response to consumption-based emissions.** Changes in annual mean geopotential height (color, m) and winds (vector) at 500 hPa (a,b) and changes in annual mean sea-level pressure (color, hPa) and winds (vector) at 850 hPa (c,d) due to $E_c$ of developing and developed countries.




**Extended Data Fig. 8 | Meridional circulation response to consumption-based emissions.** Zonal average annual mean changes in vertical velocity ($\omega = dp/dt$) due to $E_c$ of developing and developed countries. Negative values represent rising air.

**Extended Data Table 1 |** Radiative kernel decomposition of Ec induced ERF. Factors contributing to the annual mean ERF include aerosol direct radiative forcing and fast adjustments of water vapor, surface albedo, temperature and clouds. Units are in W/m2. The values in brackets represent 2 standard deviations

| | Due to $E_c$ of Developing countries | Due to $E_c$ of Developed countries |
|---|---|---|
| **ERF** | -0.48 (±0.64) | -0.42 (±0.64) |
| **Direct radiative forcing** | -0.16 (±0.031) | -0.066 (±0.026) |
| **$H_2O$** | 0.036 (±0.19) | 0.010 (±0.19) |
| **Albedo** | -0.022 (±0.16) | -0.040 (±0.17) |
| **Temperature** | -0.019 (±0.39) | -0.007 (±0.41) |
| **Cloud** | -0.32 (±0.70) | -0.32 (±0.72) |

**Extended Data Table 2 |** Radiative kernel decomposition of Ec induced TOA net radiation flux in the fully coupled simulations. Factors contributing to the annual mean net radiation flux include ERF and the responses of water vapor, surface albedo, temperature and clouds. The contributions of all non-ERF factors are in addition to the radiative effects of their fast adjustments presented in Extended Data Table 1. Units are in W/m2. The values in brackets represent 2 standard deviations

| | Due to $E_c$ of Developing countries | Due to $E_c$ of Developed countries |
|---|---|---|
| **TOA net radiation flux** | -0.073 (±0.50) | -0.12 (±0.55) |
| **ERF** | -0.48 (±0.64) | -0.42 (±0.64) |
| **$H_2O$** | -0.26 (±0.94) | -0.32 (±1.20) |
| **Albedo** | -0.046 (±0.13) | -0.060 (±0.13) |
| **Temperature** | 0.62 (±1.54) | 0.61 (±1.91) |
| **Cloud** | 0.090 (±1.05) | 0.062 (±1.23) |

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Short Communication

# The significant roles of anthropogenic aerosols on surface temperature under carbon neutrality

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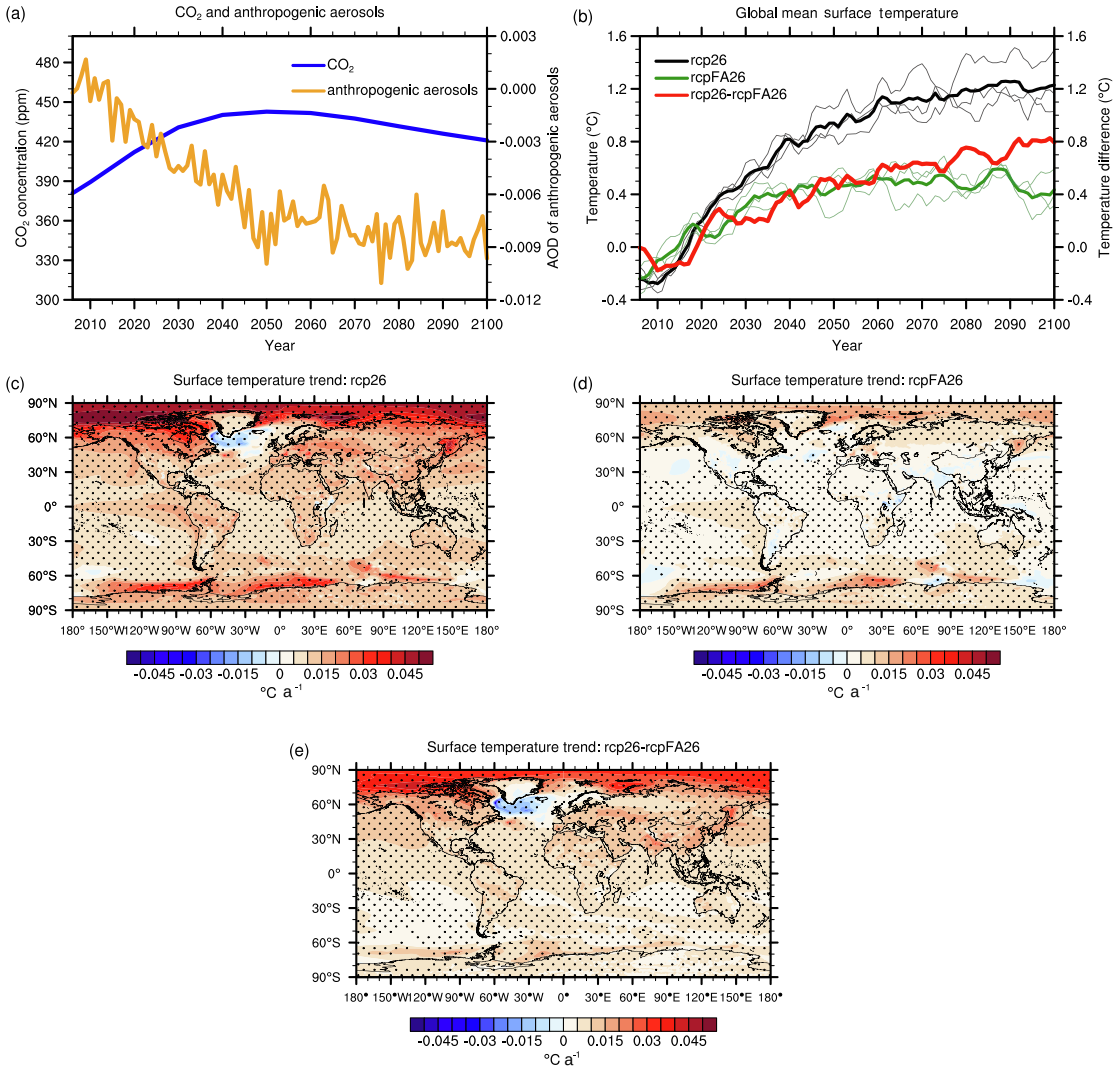
A target of limiting global warming below 1.5 or 2 °C by 2100 relative to the preindustrial level was established in the *2015 Paris Agreement* to combat the climate crisis. The fast increase in human-induced $CO_2$ in the atmosphere has accelerated the warming during the past decades. To achieve the low-warming target, China has announced that it will endeavor to reach the carbon emission peak by 2030 and carbon neutrality by 2060 [1]. Many other regions or countries have also issued legislation or policies to accomplish carbon neutrality by the middle of this century such as the European Union, India, Canada, South Africa, etc. [2]. The achievement of the ambitious carbon neutrality in the future will require anthropogenic emissions to decrease quickly from now, which could lead to reductions in both $CO_2$ and aerosols [3]. However, the same trends in $CO_2$ and aerosols have the opposite radiative forcings. The cooling effect of declined $CO_2$ will be superimposed by the warming effect of declined aerosols. Meanwhile, aerosol changes could affect oceanic processes from the surface to the depth [4,5] and alter the regional pattern of surface temperature. Therefore, the roles of anthropogenic aerosols on surface temperature changes under carbon neutrality need further investigations.

In this study, we adopt an extensively applied climate model, the Community Earth System Model version 1 with the Community Atmospheric Model version 5 [6] (CESM1-CAM5; Supplementary materials), which could well simulate the observed warming over the 20th century [6]. This model contains aerosol direct and indirect effects with the inclusion of both the cloud albedo and cloud lifetime effects. The aerosol radiative forcing in CESM1-CAM5 is −1.5 W $m^{-2}$ and within the uncertainty range at a 90% confidence level (−2.0 to −0.4 W $m^{-2}$) of models in the Coupled Model Intercomparison Project Phase 5 (CMIP5) [7]. We use simulations under the Representative Concentration Pathways 2.6 (RCP2.6), which is the only scenario near the low-warming target in CMIP5 future projections [3]. According to the released all-forcing simulations under RCP2.6 (rcp26), we conduct fixed-aerosol simulations (rcpFA26) over the 21st century in which anthropogenic aerosol emissions are fixed at the 2005 levels (Supplementary materials). The climate effects of the projected decrease in anthropogenic aerosols could be derived from the difference between these two simulations (rcp26-rcpFA26).

Under RCP2.6, the anthropogenic aerosols decrease since the beginning of the 21st century as shown by the change of aerosol optical depth (Fig. 1a), after a large reduction in anthropogenic emissions especially $SO_2$ [3]. However, $CO_2$ concentration increases before it starts to decrease after 2050 (Fig. 1a) due to the difficulties in carbon reduction technologies that could not be tackled. In rcp26, the increased $CO_2$ and decreased aerosols shall generate consistent warming effects and cause a large increase in global mean surface temperature before 2050 (0.031 °C $a^{-1}$; Fig. 1b, black). After 2050, despite the decline in $CO_2$, the decrease in anthropogenic aerosols will continue to produce a warming effect, which will enable the surface temperature to rise with a speed slower than that before 2050 (0.006 °C $a^{-1}$; Fig. 1b, black). However, in rcpFA26, the surface temperature is primarily driven by $CO_2$ and shows a slight cooling trend (−0.002 °C $a^{-1}$) after peaking at 2060 (Fig. 1b, green) without the extra warming from anthropogenic aerosols. The 10-year delay of temperature changes in response to $CO_2$ forcing is consistent with the previous studies [8]. The persistent net warming effect of anthropogenic aerosols

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**Fig. 1.** Changes in atmospheric $CO_2$ and anthropogenic aerosols as well as the response of surface temperature under RCP2.6 in CESM1-CAM5. (a) $CO_2$ concentration (blue) and aerosol optical depth (AOD) at 500 nm of anthropogenic aerosols (orange) in the 21st century. (b) Time series of global mean surface temperature anomalies (relative to 2006–2025 mean) in rcp26 (black), rcpFA26 (green), and their difference rcp26-rcpFA26 (red). Thin lines denote the results in three ensemble members. Long-term trends of surface temperature over 2006–2100 in (c) rcp26, (d) rcpFA26, and (e) their difference rcp26-rcpFA26. Trends in (c–e) are the results of the ensemble mean. Stippling indicates where SST trends are statistically significant at the 95% confidence level based on the Student's *t*-test.

leads to an increase in the surface temperature throughout the entire century (0.010 °C $a^{-1}$) as demonstrated by the simulation difference rcp26-rcpFA26 (Fig. 1b, red). The long-term evolution in the time series of global mean surface temperature is consistent with the temperature trend over 2006–2100 in spatial maps (Fig. 1c–e). The aerosol forcing makes more contributions to the total changes in the surface temperature relative to other forcings (mainly $CO_2$). This is evident in the Northern Hemisphere, especially Asia, Europe, and North America, which are the major sources of aerosol emissions (Fig. S1 online). The surface temperature does not vary consistently over the globe. The greatest warming occurs over the Arctic (Fig. 1c). In the case of declining aerosols, the Arctic temperature amplification is primarily due to the strong positive feedback between sea-ice-albedo and temperature [9] as well as an increase in horizontal energy transported to the Arctic region [10]. In contrast, a significant surface cooling dominated by the anthropogenic aerosol effects appears in the subpolar North Atlantic (Fig. 1c, e). This distinctive cooling occurring in a warming environment has been referred to as the North Atlantic warming hole and is related to variations in deep ocean circulations in the Atlantic basin [11], which is mainly the Atlantic Meridional Overturning Circulation (AMOC).

Under the net effect of decreased anthropogenic aerosols, the long-term (2006–2100) surface cooling trend over the subpolar North Atlantic (Fig. 1e) mostly comes from the second half-century, as indicated by the times series of the regionally averaged (50°–70°N, 60°–10°W) sea surface temperature (SST) in rcp26-rcpFA26 (Fig. 2c, red). The decreasing anthropogenic aerosols trigger a weakening AMOC at the beginning of the century (Figs. 2a (red) and S2 online). The initial weakening of the AMOC is possibly related to changes in atmospheric circulation [4]. The decline of aerosols reduces sea level pressure gradient, weakens surface winds over the North Atlantic, reduces surface latent and sensible heat fluxes, and ultimately leads to a decrease in surface density flux [4]. In turn, the deep convection weakens and the AMOC slows down. The weakened AMOC induces a reduction in northward ocean heat transport in the Atlantic (Fig. 2b, red). Less heat is transported from the tropics to high latitudes. As time goes by, a large SST cooling trend over the subpolar North Atlantic will emerge after 2050 (Fig. 2c, red). The SST fluctuations in the first half-cen-

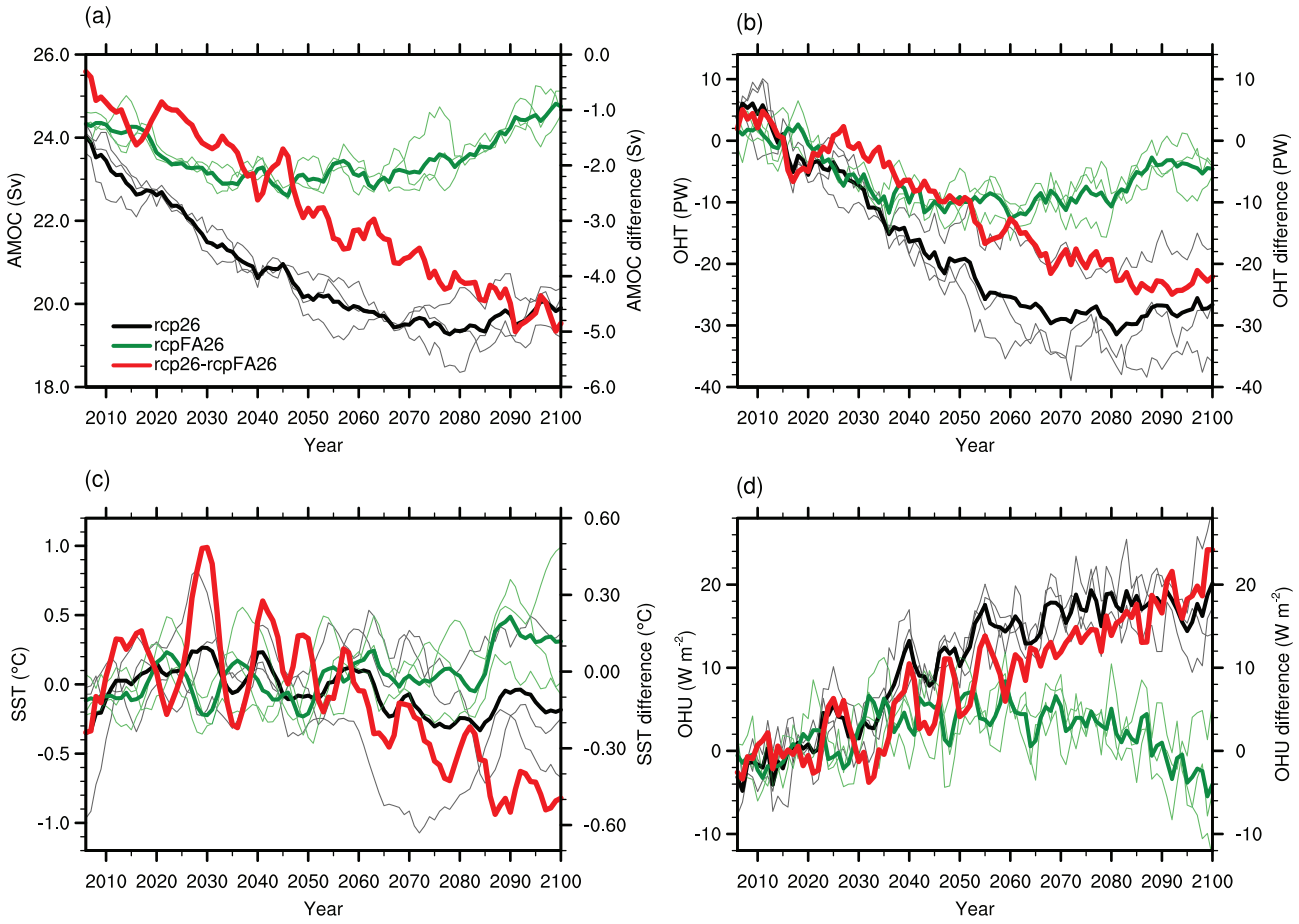



**Fig. 2.** Evolutions of oceanic processes in the Atlantic under RCP2.6 in CESM1-CAM5. Time series of (a) the AMOC index, (b) total ocean heat transport (OHT) anomalies in the Atlantic, (c) SST anomalies, and (d) ocean heat uptake (OHU) anomalies in the subpolar North Atlantic (50°–70°N, 60°–10°W) in rcp26 (black), rcpFA26 (green), and their difference rcp26-rcpFA26 (red; right *Y*-axis). Thin lines denote the results in three ensemble members. Thick lines denote the results of the ensemble mean. Anomalies are relative to the 2006–2025 mean.

tury might be associated with the competing roles between the direct warming effect of decreased aerosols and the AMOC-induced heat loss. In the second half of the century, the direct warming weakens as the aerosol reduction slows down (Fig. 1a) and the continuous weakening of the AMOC dominates the SST decline in the subpolar North Atlantic. The continuous AMOC weakening after 2050 under declining aerosols mainly results from a continuous increase of surface freshwater flux in the subpolar North Atlantic [5]. The surface freshwater flux increase is possibly controlled by a reduction in surface evaporation due to the cooling of local SST [5]. The SST decline further causes an increase in ocean heat uptake on the surface through the turbulent heat flux feedback [12] in the subpolar North Atlantic (Fig. 2d, red). The variations of ocean heat uptake in the first half of the century might be a comprehensive result of oscillating SST and direct warming effects of aerosols.

Comparing the evolutions in oceanic processes during the 21st century in rcp26 and rcpFA26 simulations (Fig. 2, black and green), we find that the extra warming effect of reduced anthropogenic aerosols in rcp26 causes a prolonged weakening in the AMOC strength (Fig. 2a, black). This subsequently leads to prolonged decreases in the Atlantic's poleward heat transport and subpolar North Atlantic SST, which in turn causes a prolonged increase in the local ocean heat uptake (Fig. 2b–d, black). However, these variables in rcpFA26 tend to recover in the second half-century, mainly following the $CO_2$ pathway (Fig. 2a–d, green). In addition to the reduced aerosol forcing, the North Atlantic warming hole could appear under increased $CO_2$ [13] via inducing a largely weakened AMOC. The warming hole in rcpFA26 is not obvious (Fig. 1d) due to the recovery of the AMOC (Fig. 2a) dominated by $CO_2$ changes.

In summary, the concentrations of $CO_2$ and aerosols in the atmosphere will decrease at the same time because anthropogenic emissions are significantly cut down to fulfill the carbon neutrality and low-warming target. The extra warming effect produced by the reduction in anthropogenic aerosols synchronizes with the cooling effect induced by $CO_2$ decrease. Thus, the reduced aerosols will allow global surface temperature to keep rising for an unexpected long period rather than gradually decline as the $CO_2$ decreases. In the 21st century, compared with the consistent long-term warming trends elsewhere, there is a long-term cooling trend of surface temperature in the subpolar North Atlantic, which leads to a large increase in local ocean heat uptake. This is primarily related to a weakening in the AMOC and a decrease in northward ocean heat transports throughout the century under the forcing of declining aerosols, which causes a substantial cooling in SST over the subpolar North Atlantic in the last decades of the century.

The response in oceans to changes in aerosol forcing could last long (e.g., decades or centuries) due to the slow adjustments of oceanic processes. In addition to modulating long-term climate change, changes in oceans could also affect the uptake and storage of atmospheric $CO_2$, which would influence carbon neutrality [1] over a long period. The North Atlantic is a major atmospheric $CO_2$ sink because the deep convection branch of the AMOC [14] leads to the deep storage of carbon, thus a weakened AMOC would

result in increased carbon concentration in the atmosphere. Beyond the CESM1 simulations under RCP2.6 scenario, the latest CESM2 model and the updated low-emission scenario SSP1-2.6 in the framework of CMIP6 could be applied in the next step of the study. One caveat in this study is the potential model dependence of the aerosol-induced climate responses. The current climate models differ in aerosol schemes [7], deep convection sites [15], parameterization, etc. The findings from the CESM model need to be compared with those in other climate models.

Our study indicates that future decreases in anthropogenic aerosols could amplify the warming maximum and slow down the cooling rate due to $CO_2$ reduction, thereby extending the time required to reach the low-warming target. Despite the 1.5 °C/2 °C temperature goal set in the *Paris Agreement*, the zero-emission pathways in practice are subject to various possibilities. The specific future emissions are different among countries due to their different policies [2]. The carbon emission peak and emission reduction rate could be distinct from those under RCP2.6. Hence, future climate change under carbon neutrality could have some biases against that projected under RCP2.6. Nevertheless, the significant roles played by anthropogenic aerosols on the spatiotemporal evolution of surface temperature need to be considered in designing the specific pathway of carbon neutrality.

## Conflict of interest

The authors declare that they have no conflict of interest.

## Acknowledgments

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## Author contributions

Gang Huang and Xiaofan Ma conceived the research. Xiaofan Ma performed model simulations, prepared figures, and led the writing of the manuscript. Junji Cao contributed to the discussion and explanation of the results.

## Appendix A. Supplementary materials

Supplementary materials to this article can be found online at https://doi.org/10.1016/j.scib.2021.10.022.

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OCEANOGRAPHY

# Greenhouse warming intensifies north tropical Atlantic climate variability

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**Variability of North Tropical Atlantic (NTA) sea surface temperature (SST), characterized by a near-uniform warming at its positive phase, is a consequential mode of climate variability. Modulated by El Niño–Southern Oscillation (ENSO) and the North Atlantic Oscillation, NTA warm anomalies tend to induce La Niña events, droughts in Northeast Brazil, increased frequency of extreme hurricanes, and phytoplankton blooms in the Guinea Dome. Future changes of NTA variability could have profound socioeconomic impacts yet remain unknown. Here, we reveal a robust intensification of NTA variability under greenhouse warming. This intensification mainly arises from strengthening of ENSO-forced Pacific-North American pattern and tropospheric temperature anomalies, as a consequence of an eastward shift of ENSO-induced equatorial Pacific convection and of increased ENSO variability, which enhances ENSO influence by reinforcing the associated wind and moist convection anomalies. The intensification of NTA SST variability suggests increased occurrences of extreme NTA events, with far-reaching ramifications.**

## INTRODUCTION

Variability of North Tropical Atlantic (NTA) sea surface temperature (SST) (hereafter the NTA) typically peaks in boreal spring (March, April, and May, MAM1, where "1" refers to the current year) and is characterized by basin-wide SST warming at its positive phase (*1*, *2*). Coupled with a latitudinal movement of the intertropical convergence zone, the NTA profoundly influences precipitation in Northeast Brazil and Sahel (*3*–*6*). An anomalous NTA warming during 1979 to 1981 induced severe drought in Northeast Brazil, leading to a more than 70% reduction in production of rice, beans, and cotton (*7*). An intense NTA warming contributed to the 2012 to 2016 drought in Northeast Brazil, affecting 33.4 million people and resulting in losses of U.S. $30 billion (*8*). By modulating the intensity, number, and track pattern of Atlantic tropical cyclones, an NTA warm event increases the number of major hurricanes and their landfall frequency along the U.S. East Coast (*9*–*12*). Moreover, the NTA has a prominent influence on the Guinea Dome (*13*), chlorophyll-α concentration, and ecosystems (*14*, *15*). In addition to regional influences, the NTA exerts its climatic impacts over the globe, including El Niño–Southern Oscillation (ENSO) in the Pacific (*16*), sea ice distribution in the Antarctic (*17*), and anomalous global mean temperature (*18*). Because of these severe effects, determining the response of NTA to greenhouse warming is an issue of great importance.

The NTA mainly arises from latent heat flux anomalies associated with anomalous northeasterly trades (*19*–*21*). Specifically, the wind-evaporation-SST feedback (*22*), mainly confined to the deep tropics, contributes to the development of anomalous SST (*23*, *24*), but forcings outside of the tropical Atlantic are required to reinforce the NTA temperature anomaly (*2*, *25*, *26*). The North Atlantic Oscillation (NAO) is one such forcing (*27*, *28*). During a negative NAO event, the northeasterly trades weaken in response to a slackened Subtropical High, reducing surface latent heat flux and leading to an anomalous NTA warming. Another important forcing comes from ENSO (*29*, *30*), which exerts its influence in several pathways. During El Niño, tropospheric temperature over the central and eastern equatorial Pacific increases and propagates eastward in the form of equatorial Kelvin waves, which reduces moist convection over the northern tropical Atlantic and gives rise to warm SST anomalies, a process referred to as "tropospheric temperature" mechanism (*20*, *31*). Furthermore, convective anomalies associated with increased precipitation in the western and central equatorial Pacific excite the Pacific–North American (PNA) pattern (*32*), a Rossby wave train with a ridge over the western North America and troughs over the Aleutians and the southeastern United States. In response, the northeasterly trades weaken over the northern tropical Atlantic, reducing evaporation and generating warm SST anomalies. In addition, El Niño induces a negative diabatic heating over the Amazon basin, which, in turn, generates an anomalous Atlantic Hadley circulation (*33*) and a Gill-type response (*34*), contributing to the weakened northeasterly trades.

Despite the advances described above, how the NTA may respond to greenhouse warming remains unknown. Below, we show that most of the Coupled Model Intercomparison Project Phase 6 (CMIP6) models simulate an increase in NTA variability.

## RESULTS

### Observed NTA SST variability

To identify the observed NTA, we applied empirical orthogonal function (EOF) analysis to quadratically detrended SST anomalies (see the "EOF analysis" section in Materials and Methods) averaged

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over boreal spring when the NTA reaches its peak. Here, we take the normalized principal component (PC) time series as the NTA index. Given that ENSO can trigger NAO events (*35*, *36*), a partial regression (partial correlation) is conducted to extract the "pure" NAO influence on (connection with) the NTA (see the "Partial regression" and "Partial correlation" sections in Materials and Methods).

As the leading mode in boreal spring, the observed NTA accounts for 64% of the total SST variance (fig. S1). It is tightly related with ENSO and the NAO in previous winter (December, January, and February, D0JF1, where "0" refers to the previous year), due to their influence on northeasterly trades and latent heat flux (Fig. 1, A and B). The (partial) correlation coefficient between MAM1 NTA and D0JF1 ENSO (NAO) reaches 0.59 (−0.5) over the period of 1951 to 2014 (red star, Fig. 1C).

## Model selection

We conducted similar EOF analysis using outputs from 23 CMIP6 models that are forced with historical forcing during the same period (1951 to 2014) as in observations. EOF analysis is useful in extracting the pattern of NTA variability, allowing models to have different spatial patterns. The NTA emerges as the first EOF mode in each model, consistent with the observations. However, several models underestimate the NTA connections with ENSO or the NAO (Fig. 1C). Among them, 16 models produce reasonable connections, with both correlation coefficients exceeding 50% in amplitude of the observed values. These models reproduce reasonably well the observed influences from ENSO and the NAO (Fig. 1, A and B versus D and E), with maximum wind and SST anomalies located over the NTA center (box in Fig. 1), although in most models, the influence is underestimated for both ENSO and the NAO. By comparison, the remaining seven models, featuring a westward (northward) shift of anomalous trades, generate weaker SST and wind responses to ENSO (the NAO) forcing (Fig. 1, A and B versus F and G). These wind shifts are related to a westward-displaced ENSO center (*37*, *38*) and an underestimated magnitude of NAO variability simulated in models (*2*, *39*). Nevertheless, the selected models reasonably simulate the amplitude of observed NTA variability (0.40°C, defined by SST SD averaged over the NTA center), with a range from 0.30° to 0.46°C and an ensemble mean of 0.35°C. Therefore, we use the 16 models that realistically simulate the connections with ENSO and the NAO to study the response of the NTA to greenhouse warming.

## Increased NTA variability in the future climate

We performed EOF analysis to quadratically detrended MAM1 SST anomalies over the period of 1900 to 2099 in the 16 selected models. These models are forced with historical forcing until 2014 and with Shared Socioeconomic Pathways (SSP) 2-4.5 emission scenario from 2015 onward. We compared the SD of NTA variability between the present (1900 to 1999) and future (2000 to 2099) climate. A total of 12 of the 16, or 75%, selected models simulate increased NTA variability (Fig. 2A). The ensemble mean increase is 8.3%, significant above the 95% confidence level according to a bootstrap test (see the "Bootstrap test" section in Materials and Methods) (fig. S2A). This increase is underpinned by a robust (15 of the 16 selected models) enhancement of SST variability averaged over the NTA center (fig. S3A), with a significant ensemble mean increase of 10% (fig. S3B).

In association, occurrences of extreme NTA events, defined as |PC| greater than 1.75 SD, become more frequent under greenhouse warming (Fig. 2B). Aggregated over the 16 models, the frequency increases by 38%, from 1 event every 14.7 years (109 events in 1600 years) in the present period to 1 event every 10.7 years (150 events in 1600 years) in the future period. This increase in frequency enjoys a strong intermodel consensus (Fig. 2B) and an ensemble mean increase significant above the 95% confidence level (fig. S2B). In addition, the frequency of extreme events increases by 21 or 57%, when using threshold values of 1.5 or 2.0 SD, respectively, further supporting our result.

A sensitivity test further reveals that the increase in NTA variability is not sensitive to model selection. Even when all the 23 models are considered, the intermodel consensus on the increased NTA variability is still significant, with 16 of the 23 models (70%) producing an increase (fig. S4A). Identically, the increased frequency of extreme NTA events is still robust when considering all the 23 models (fig. S4B).

## Enhancement in ENSO influence under greenhouse warming

An ensemble mean of the NTA response to NAO forcing barely changes in the future climate, indicating a minor role of the NAO in driving increased NTA variability (fig. S5). Instead, a change in ENSO teleconnection and in ENSO variability under greenhouse warming is the key driver of the increase in variability of the NTA.

To describe ENSO, we use Niño3.4 index normalized by the SD over the full 200 years to enhance intermodel comparability. Total response of the NTA to ENSO forcing in each period is calculated by regressing the NTA onto the Niño3.4 index and then multiplying the regression coefficient by 1 SD value of the Niño3.4 index. A comparison of the total response shows a substantial enhancement in a warming climate (Fig. 3A) (see the "Regression analysis" section in Materials and Methods). The enhanced response to ENSO is found consistently in all the 16 models except one (94% of the models), with an ensemble mean enhancement of 20% (from 0.56 in the present climate to 0.67 in the future climate) statistically significant according to a bootstrap test (Fig. 3B). In association, ENSO influence strengthens (Fig. 3, C and D), contributing to the increase in NTA variability. Specifically, models with a greater enhancement of NTA response to ENSO forcing produce a larger increase in NTA variability, and the intermodel correlation between changes in NTA response to ENSO and in NTA variability reaches 0.83 (Fig. 3E). Moreover, the enhanced NTA response to ENSO forcing is related to more frequent extreme NTA events, while that to the NAO forcing barely contributes (fig. S6).

### *An eastward intensification of ENSO teleconnection*

The enhanced NTA response to ENSO forcing is, in part, induced by an eastward shift of ENSO-induced response in convection (Fig. 4A). In the present climate, precipitation response to ENSO is centered over the western equatorial Pacific (fig. S7A). Under greenhouse warming, these anomalies intensify in the central to eastern equatorial Pacific (black box, fig. S7B), suggesting an eastward shift of convective anomalies (*40*–*44*). This shift in convection strengthens the sensitivity of PNA teleconnection to ENSO forcing (*40*), especially over the lobe of southeast United States, as seen in sensitivity of 200-hPa geopotential height anomalies to the Niño3.4 index (fig. S8). Specifically, 14 of 16 (88%) models produce a deepening of low-pressure anomalies, with an ensemble mean increase of 39%.

### *Increased ENSO variability*

In addition, ENSO variability increases in a warming climate, enhancing its remote forcing on NTA variability (Fig. 4B). A total of

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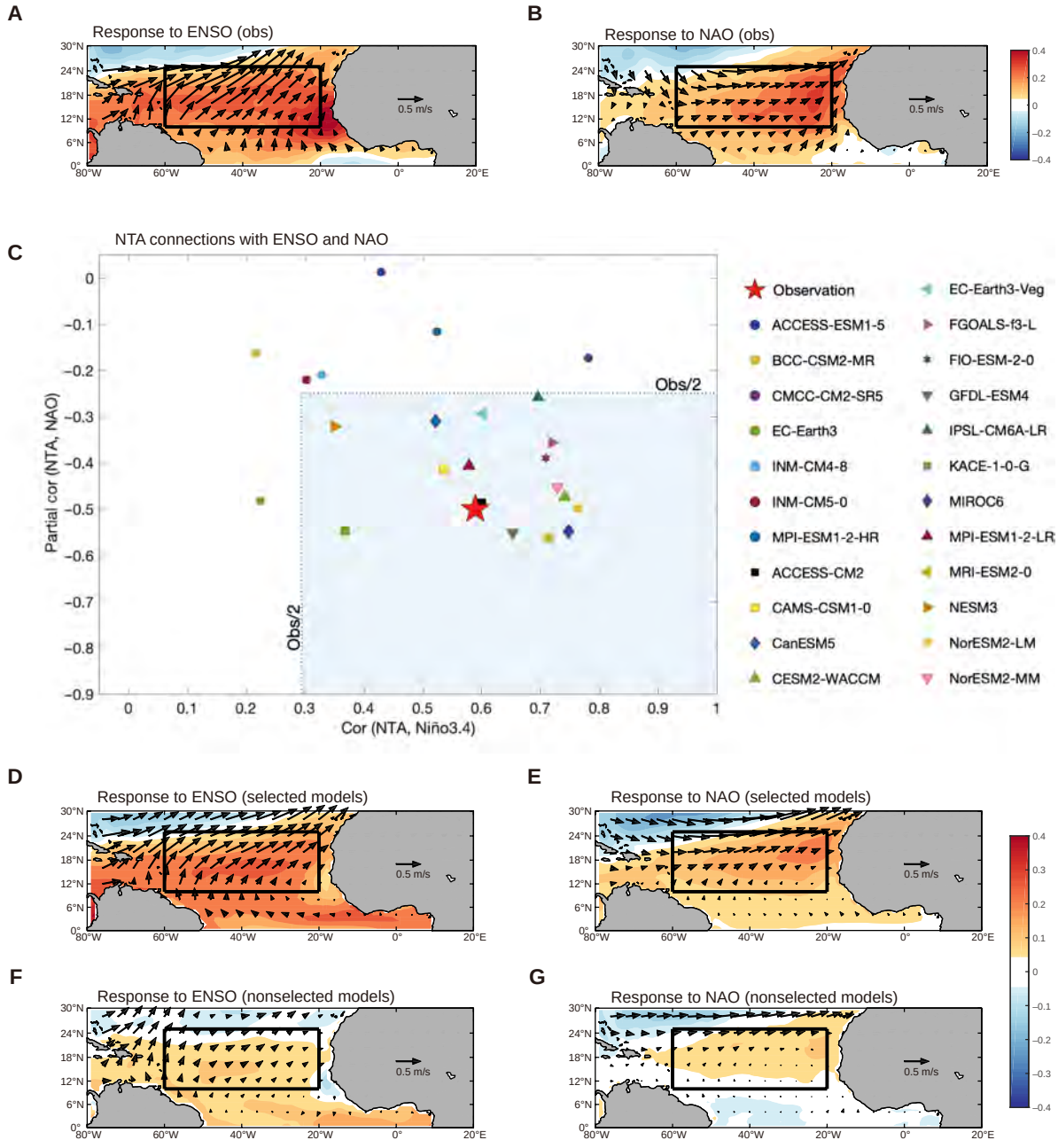



**Fig. 1. NTA connections with ENSO and the NAO in observation and CMIP6 models.** (**A**) Response of MAM1 SST (°C, color shading) and FMA1 wind (meters per second, vector) to observed Niño3.4 over the period of 1951 to 2014, calculated by regressing SST and wind anomalies onto Niño3.4 index and then multiplying the regression coefficients by 1 SD value of Niño3.4 index. (**B**) Same as (A) except for partial regression against an NAO index (reversed sign) by removing variability associated with Niño3.4 index (see the "Partial regression" and "EOF analysis" sections in Materials and Methods for NTA, NAO, and Niño3.4 indices). (**C**) Correlation coefficients between the NTA and Niño3.4 (*x* axis) and partial correlation coefficients between the NTA and the NAO (*y* axis; see the "Partial correlation" section in Materials and Methods). The large red-filled star indicates the observed value. In seven models, marked by circles, two generate an NTA-ENSO correlation and six produce an NTA-NAO correlation that are smaller than 50% in amplitude of the observations. These are referred as nonselected models. The remaining 16 models in blue quadrant are selected models. The dashed lines represent 50% of the observed values. (**D** and **E**) Same as (A) and (B) but for an ensemble mean of 16 selected models. (**F** and **G**) Same as (A) and (B) but for an average of the two models that produce a too-weak connection between the NTA and ENSO (correlation coefficients lower than 50% of the observed) (F) and for an ensemble mean of the six models that fail to generate the connection with the NAO (G). The boxes represent the NTA center (10°N to 25°N and 60°W to 20°W).

12 of the 16 (75%) models produce increased Niño3.4 variability, with an ensemble mean increase of 11% (fig. S9). When all 23 models are considered, an increase in variability is simulated by 16 of the 23 (70%) models (fig. S4C), consistent with previous findings of enhanced variability in both eastern Pacific and central Pacific El Niño (*45*). The enhancement in ENSO variability further strengthens the ENSO-forced response in PNA teleconnection. Here, we examine the total change in the PNA teleconnection that includes change in PNA

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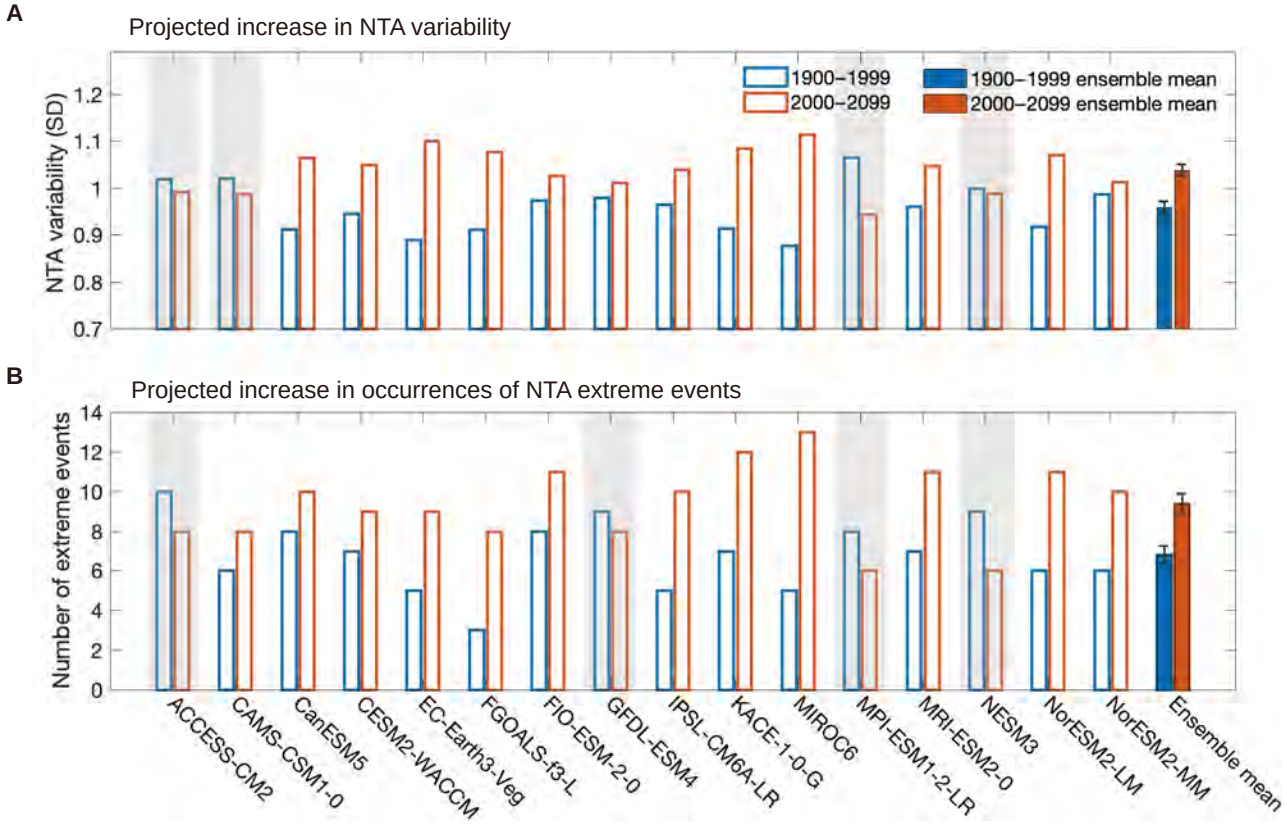



**Fig. 2. Projected increase in NTA variability and occurrences of extreme NTA events.** (**A**) Comparison of the NTA SD over the present (1900 to 1999, blue-edged bars) and future (2000 to 2099, red-edged bars) periods in the 16 selected models. (**B**) Same as (A), but for the number of extreme events (|PC| > 1.75 SD). The multimodel ensemble means over the present and future periods are shown in blue-filled and red-filled bars, respectively; error bars are calculated as 1 SD of a total of 10,000 interrealizations of a bootstrap method (see the "Bootstrap test" section in Materials and Methods); models that simulate a reduction are grayed out. Variability of the NTA and the frequency of extreme NTA events are projected to increase under greenhouse warming.

sensitivity to ENSO and change in ENSO variability itself. Hence, for each individual period, we regress geopotential height anomalies onto the Niño3.4 index and then multiply the regression coefficients by 1 SD value of the Niño3.4 index. In this case, a deepening of the low-pressure anomalies over the southeast United States is simulated by 15 of the 16 models (94%), with an ensemble increase reaching 53% (Fig. 4, C to E). The strengthening of PNA teleconnection is also seen in sea level pressure anomalies (fig. S10).

***Consequence of enhanced ENSO teleconnections***

As a result of the intensification of the PNA pattern, ENSO-forced response in wind is enhanced over the northern tropical Atlantic (Fig. 3, C and D), with a maximum increase north of 15°N. The enhanced wind speed response, with an ensemble mean increase of 21%, significantly contributes to the increase in the NTA response to ENSO forcing (Fig. 5A). The intermodel correlation between changes in the wind response and in the NTA response reaches −0.83.

In addition, SST of the NTA center becomes more sensitive to wind forcing, as seen in MAM1 SST anomalies regressed onto unnormalized February, March, April (FMA1) wind speed anomalies (fig. S11A), which measures the sensitivity of SST to 1 m/s of wind speed change. In a warming climate, the ocean mixed layer depth shoals over the northern tropical Atlantic, a robust feature among all models (using all models with available outputs; fig. S11B). This shoaling arises from an intensification of oceanic stratification (*46*–*48*) and from a weakening of climatological northeasterly trades (using all models with available outputs; fig. S11, C and D). As a result of the shallower mixed layer, SST response to a given wind forcing increases in most of the models. However, its contribution to the NTA response to ENSO forcing is not statistically significant and therefore plays a minor role (fig. S11E).

Furthermore, under greenhouse warming, tropospheric temperature anomalies are intensified due to increased ENSO variability (fig. S12) and an amplified vapor response to ENSO (*49*), contributing to the enhanced NTA response to ENSO forcing (Fig. 5B). An intermodel correlation between changes in ENSO-forced response of tropospheric temperature and changes in the NTA response to ENSO forcing ($R = 0.63$), although not as high as that between changes in ENSO-forced response of wind speed and changes in the NTA response to ENSO forcing ($R = -0.83$; Fig. 5A), indicates a substantial contribution from the tropospheric temperature mechanism.

## DISCUSSION

ENSO-forced responses in PNA teleconnection and tropospheric temperature are important mechanisms in driving the variability of NTA (Fig. 6A). Both of these mechanisms enhance under greenhouse warming, due to an eastward shift of ENSO-induced response in equatorial Pacific convection and an enhancement of ENSO

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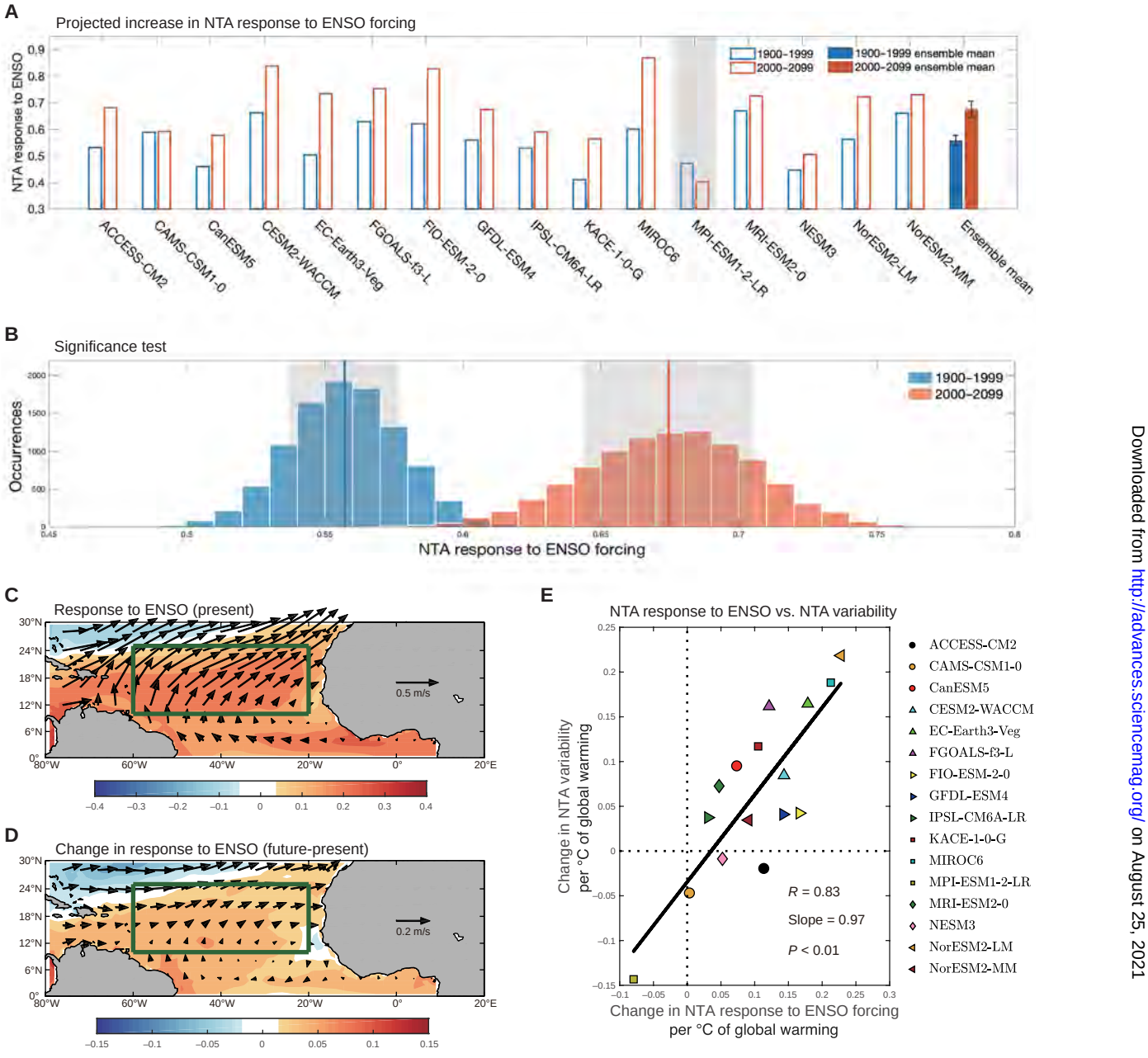



**Fig. 3. Change in NTA response to ENSO forcing and the projected increase in NTA variability.** (**A**) Comparison of NTA response to ENSO forcing (SD) over the present (1900 to 1999, blue-edged bars) and future (2000 to 2099, red-edged bars) periods in the 16 selected models. The multimodel ensemble means over the present and future periods are shown in blue-filled and red-filled bars, respectively; error bars are calculated as 1 SD of a total of 10,000 inter-realizations of a bootstrap method (see the "Bootstrap test" section in Materials and Methods); models that simulate a reduction are grayed out. (**B**) Histograms of 10,000 realizations of a bootstrap method for NTA response to ENSO forcing in the present (1900 to 1999, blue) and future (2000 to 2099, red) climate. The blue and red lines indicate the mean values of the 10,000 realizations for the present and future periods, respectively. The gray shaded areas refer to the respective 1 SD of the 10,000 realizations. (**C** and **D**) Response of MAM1 SST (°C, color shading) and FMA1 surface winds to ENSO (meters per second, vector), in the present climate (C) and their projected changes (future minus present) (D). Green boxes (10°N to 25°N, 60°W to 20°W) represent the NTA center. (**E**) Intermodel relationship between the change in NTA response to ENSO (SD) and in NTA variability (SD). To enhance intermodel comparability, we scale the changes by an increase in global mean SST of each model. Linear fit (black solid line) is shown together with the correlation coefficient *R*, slope, and *P* value from the regression. The NTA response to ENSO forcing is projected to increase, contributing to enhanced NTA variability.

variability. In association, ENSO-forced responses in wind speed and tropospheric temperature strengthen under greenhouse warming (Fig. 6B), leading to the increase in the amplitude of NTA SST variability and the frequency of extreme NTA events.

We note that most of the models simulate an overly weak relationship between NTA and NAO variability. Although the weak relationship might represent a source of uncertainty, there is no intermodel consensus on a simulated increase in NAO variability

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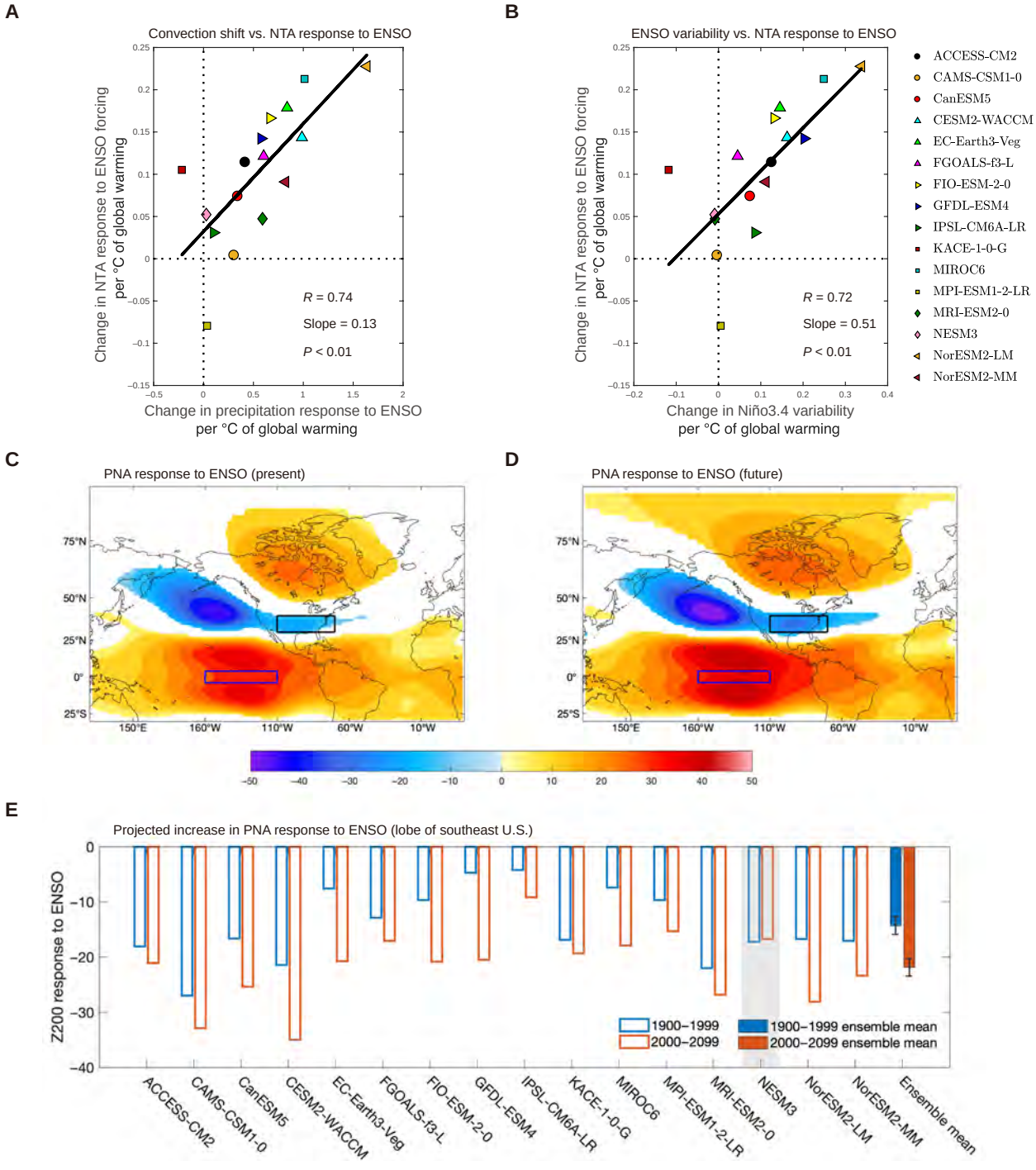



**Fig. 4. Strengthening of PNA teleconnection.** (**A** and **B**) Intermodel relationships between change in NTA response to ENSO (SD) and in FMA1 precipitation response to ENSO (millimeters per day) averaged over the central to eastern equatorial Pacific [blue boxes in (C) and (D)] (A) and in Niño3.4 variability (SD) (B). To enhance intermodel comparability, we scale the changes by an increase in global mean SST of each model. Linear fits (black solid line) are shown together with the correlation coefficients *R*, slopes, and *P* values from the regression. (**C** and **D**) Multimodel ensemble mean of the D0JF1 200-hPa geopotential height response to ENSO (meters), in the present (1900 to 1999) (C) and future (2000 to 2099) (D) climate. Only values exceeding 1 SD of intermodel spread are shown in color shading. Black and blue boxes represent the PNA center over the southeast United States (30°N to 40°N, 110°W to 70°W) and the central to eastern equatorial Pacific (4°S to 4°N, 160°W to 110°W), respectively. (**E**) Comparison of the D0JF1 200-hPa geopotential height response to ENSO (meters), averaged over the southeast United States [black boxes in (C) and (D)], in the present (1900 to 1999, blue-edged bars) and future (2000 to 2099, red-edged bars) periods in the 16 selected models. The multimodel ensemble means over the present and future periods are shown in blue-filled and red-filled bars, respectively; error bars are calculated as 1 SD of a total of 10,000 inter-realizations of a bootstrap method (see the "Bootstrap test" section in Materials and Methods); models that simulate a reduction are grayed out. The ENSO-induced PNA teleconnection is projected to increase, due to an increase in ENSO variability and an eastward shift of ENSO-induced equatorial Pacific convection.

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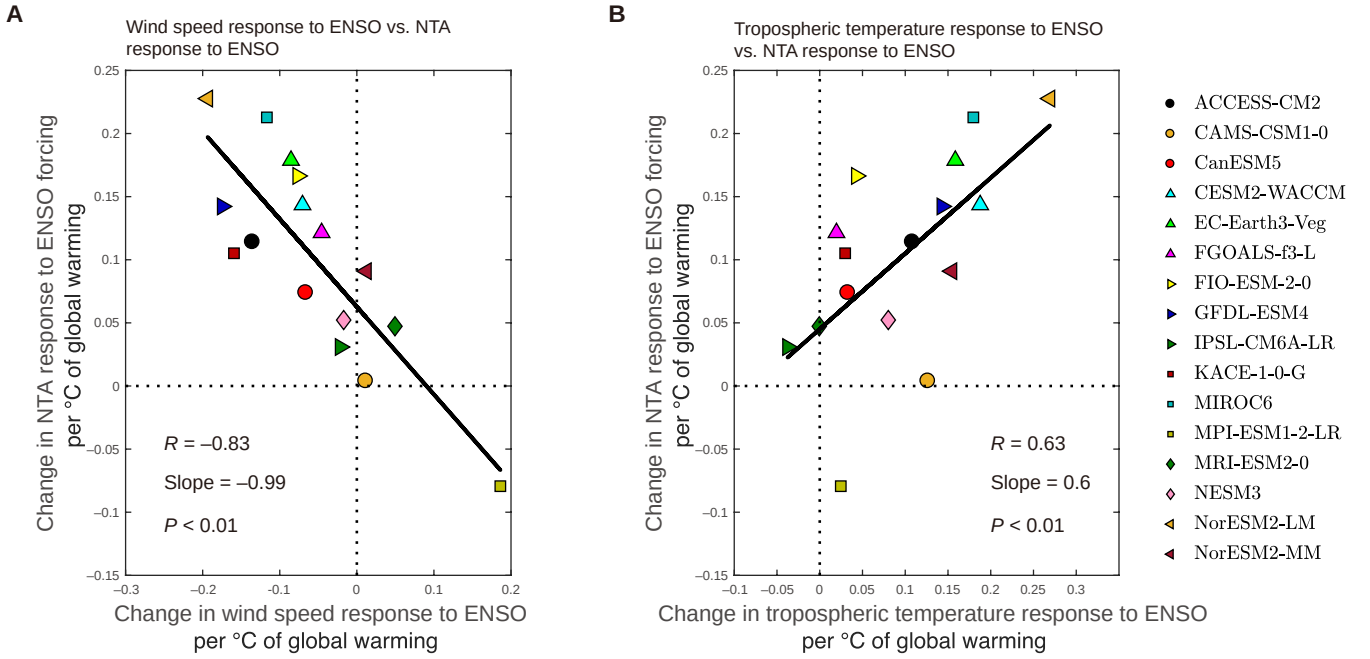



**Fig. 5. Mechanism for the projected increase in NTA response to ENSO forcing.** Intermodel relationships between change in NTA response to ENSO (SD) and in FMA1 wind speed response to ENSO (meters per second) averaged over the NTA center (**A**) and in tropospheric temperature response to ENSO (°C) averaged over the NTA center (**B**). To enhance intermodel comparability, we scale the changes by an increase in global mean SST of each model. Linear fits (black solid line) are shown together with the correlation coefficients *R*, slopes, and *P* values from the regression. The projected increase in NTA response to ENSO is primarily due to increased ENSO-induced wind speed variability and considerably due to increased ENSO-induced tropospheric temperature variability.

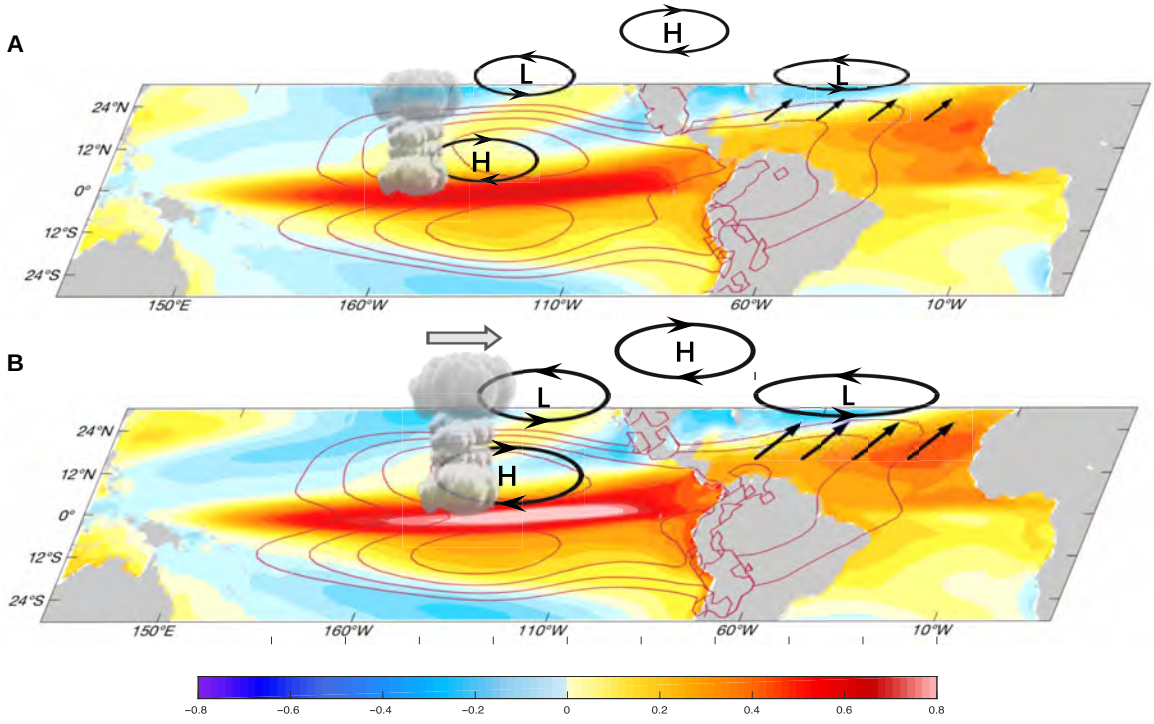



**Fig. 6. Schematic diagram of the physical mechanisms responsible for increased NTA variability.** ENSO-induced MAM1 SST (color shading), wind (vectors), tropospheric temperature (red contour), convection (cloud) anomalies, and the associated PNA (high/low-pressure anomalies) teleconnection seen in the present-day climate (**A**) intensify and undergo an eastward shift (marked by the gray arrow) as greenhouse warming proceeds into the future climate (**B**), leading to increased NTA variability in the future climate. H, high; L, low.

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(*39*), indicating an inconsequential impact from NAO variability on the increase in NTA variability. This result could change if there is a forced NAO signal that is currently underestimated (*50*). Other biases are found over the tropical Atlantic and the Pacific, for example, a bias of a double intertropical convergence zone over the Atlantic and a weaker-than-observed influence of the tropical Atlantic on ENSO (*51*–*53*). Hence, interactions between NTA variability and ENSO, and ENSO properties, may be affected by model biases. However, the nature of impacts from tropical biases on the change of tropical Atlantic SST variability is not clear, and there is no evidence that these biases may affect the strengthening of ENSO-forced responses in PNA teleconnection and tropospheric temperature, the main factors that force the increase in NTA variability.

Thus, the consequence of an increase in ENSO variability and its teleconnections under greenhouse warming is more severe than previously thought, as the increase can energize dominant modes of climate variability remote from the Pacific, such as the NTA. Given the profound climatic impacts of the NTA in inducing droughts, floods, and extreme Atlantic hurricanes in affected regions, which are projected to increase (*54*, *55*), our results add to the urgency of reducing emissions of greenhouse gases.

## MATERIALS AND METHODS

### Observed and CMIP6 data

To characterize the NTA connection with ENSO and with the NAO, we use monthly SST reanalysis from Hadley Centre Sea Ice and Sea Surface Temperature dataset version 1.1 (HadISST v1.1) (*56*) during 1951 to 2014 and monthly 200- and 500-hPa geopotential height from the National Centers for Environmental Prediction and the National Center for Atmospheric Research global reanalysis (*57*) for the period of 1951 to 2014. Also, we analyze outputs from 23 CMIP6 models, with historical and ssp2-4.5 emission scenario forcings over the period of 1900 to 2014 and 2015 to 2099, respectively (table S1) (*58*).

### EOF analysis

To capture observed and model-simulated NTA variability, EOF analysis (*59*) is performed on quadratically detrended MAM1 SST anomalies over the northern tropical Atlantic (0°N to 30°N and 80°W to 20°E). The NTA index is referred to as the normalized PC time series. To identify the NAO variability, the same approach was applied to quadratically detrended 500-hPa geopotential height over the North Atlantic (20°N to 70°N and 60°W to 0°E) in D0JF1. ENSO is represented by the normalized Niño3.4 index (SST anomalies averaged over 5°S to 5°N and 170°W to 120°W) after quadratically detrended in D0JF1. All data have been quadratically detrended before analysis.

### Regression analysis

Regression analysis is conducted to study the influence of $B$ on $A$. When regressing index $A$ onto index $B$, we are interested in the sensitivity of $A$ to $B$. When regressing $A$ onto $B$ and multiplying the slope by the SD of $B$, we are interested in the total response of $A$ to $B$, resulting from the sensitivity of $A$ to $B$ and SD of $B$.

### Bootstrap test

Bootstrap test (*60*) is conducted to examine whether a change is statistically significant. A total of 10,000 realizations are conducted to obtain the mean from the 16 selected models. Each realization is averaged over 16 samples that are independently and randomly resampled from the 16 selected models. In this resampling process, any model is allowed to be selected more than once. The SD of the 10,000 realizations is calculated for each period. If the mean value difference between the future and present periods is greater than the sum of the two separate 10,000-realization SD values, then the change is statistically significant above 95% confidence level.

### Partial correlation

To exclude ENSO influence and achieve the "pure" relationship between the NAO and NTA, a partial correlation analysis is applied

$$R_{Y(B|A)} = \frac{R_{YB} - R_{YA} * R_{AB}}{\sqrt{1 - R_{YA}{}^2} * \sqrt{1 - R_{AB}{}^2}} \quad (1)$$

where $Y$, $B$, and $A$ represent the variables (such as the NTA), the NAO index, and Niño3.4 index, respectively. $R$ denotes the correlation coefficient, and $R_{Y(B|A)}$ for partial correlation coefficient between the variable $Y$ and NAO index ($B$), after the influence of ENSO ($A$) is removed from NAO ($B$).

### Partial regression

To obtain the "pure" influence of NAO, a partial regression analysis is applied to exclude the ENSO influence. The partial regression coefficient is as

$$C_{Y(B|A)} = \frac{R_{YB} - R_{YA} * R_{AB}}{\sqrt{1 - R_{AB}{}^2}} * \frac{S_Y}{S_{B|A}} \quad (2)$$

where $C_{Y(B|A)}$ is the partial regression coefficient between the target variable $Y$ and NAO index ($B$), after the influence of ENSO ($A$) is removed from NAO ($B$). $S_Y$ and $S_{B|A}$ represent the SD of variable $Y$ and NAO ($B$) after removing the factor Niño3.4 ($A$).

## SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/7/35/eabg9690/DC1

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# Greenhouse warming intensifies north tropical Atlantic climate variability

Yun Yang, Lixin Wu, Ying Guo, Bolan Gan, Wenju Cai, Gang Huang, Xichen Li, Tao Geng, Zhao Jing, Shujun Li, Xi Liang and Shang-Ping Xie

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# Intensification of El Niño-induced atmospheric anomalies under greenhouse warming

Kaiming Hu[1,2,3] ✉, Gang Huang[1,2,4] ✉, Ping Huang[1,3], Yu Kosaka[5] and Shang-Ping Xie[6]

**The El Niño/Southern Oscillation (ENSO) has a profound influence on global climate and ecosystems. Determining how the ENSO responds to greenhouse warming is a crucial issue in climate science. Despite recent progress in understanding, the responses of important ENSO characteristics, such as air temperature and atmospheric circulation, are still unknown. Here, we use a suite of global climate model projections to show that greenhouse warming drives a robust intensification of ENSO-driven variability in boreal winter tropical upper tropospheric temperature and geopotential height, tropical humidity, subtropical jets and tropical Pacific rainfall. These robust changes are primarily due to the Clausius–Clapeyron relationship, whereby saturation vapour pressure increases nearly exponentially with increasing temperature. Therefore, the vapour response to temperature variability is larger under a warmer climate. As a result, under global warming, even if the ENSO's sea surface temperature remains unchanged, the response of tropical lower tropospheric humidity to the ENSO amplifies, which in turn results in major reorganization of atmospheric temperature, circulation and rainfall. These findings provide a novel theoretical constraint for ENSO changes and reduce uncertainty in the ENSO response to greenhouse warming.**

The El Niño/Southern Oscillation (ENSO) arises from complex interactions between the atmosphere and ocean[1–5]. On a timescale of 2–7 years, the eastern equatorial Pacific varies between anomalously cold (La Niña) and warm (El Niño) conditions. These swings in sea surface temperature (SST) shift atmospheric convection in the tropical Pacific and cause tropical and extratropical climate anomalies by changing the Walker circulation and exciting atmospheric teleconnection such as the Pacific–North American pattern[6–9]. Furthermore, the ENSO-induced tropical tropospheric temperature anomalies spread from the Pacific to all longitudes[10–12], broadening climate variability around the globe[12–15].

Because of the climatic effects around the globe, how the ENSO and its impacts might change in response to greenhouse warming has been investigated extensively during the past 20 years[16–24]. Coupled general circulation models (CGCMs) disagree on ENSO changes in SST due to the complexity of air–sea feedbacks in the tropical Pacific Ocean[16,25,26]. Therefore, understanding and determining how the ENSO responds to greenhouse warming remains a great challenge. A robust change in ENSO-driven rainfall variability was identified in CGCM projections[27]. Most CGCMs project an intensification of El Niño-driven drying over the western Pacific Ocean and wetting over the central and eastern equatorial Pacific[18,27–29]. Associated with changes in rainfall variability, the ENSO-forced Pacific–North American teleconnection pattern shifts eastward[30–32] and ENSO-driven precipitation variability is projected to increase in many regions around the globe[33]. However, it remains unclear how other important ENSO characteristics, such as tropospheric temperature, humidity and atmospheric circulation, might change under greenhouse warming.

## Tropical humidity and temperature variability

Based on output from 19 Coupled Model Intercomparison Project Phase 6 (CMIP6) models, we examined changes in the ENSO between 1% $yr^{-1}$ [$CO_2$] increase experiments and pre-industrial control experiments. We compared climate in two periods: 400–449 model years in the pre-industrial control experiments ($P_{ctrl}$) and 100–149 model years in the 1% $yr^{-1}$ [$CO_2$] experiments ($P_{warm}$). From $P_{ctrl}$ to $P_{warm}$, the multi-model ensemble (MME) mean of December-to-February (DJF) climatological tropical-mean (10° S–10° N) SST increased by 3.7 K, with the maximum increase located in the equatorial eastern Pacific (Fig. 1a), in agreement with previous studies[34,35]. Figure 1b compares the standard deviations of the DJF Niño3.4 SST index between two periods in each model and the MME. Hereafter, the linear trends in Niño3.4 SST index, specific humidity, air temperature, wind and geopotential height have been removed to isolate interannual variability. A different method[36] to detrend the data produced almost the same results (Extended Data Fig. 1). Among the 19 models, 11 showed an increase while the other eight showed a decrease in the standard deviation of the Niño3.4 SST index from $P_{ctrl}$ to $P_{warm}$, highlighting a lack of consensus in the projection of SST variability, consistent with previous studies[16]. Nevertheless, the ENSO-driven variability in DJF tropical-mean (10° S–10° N) surface air specific humidity ($q'_{sfc}$) showed a coherent increase in every model (Fig. 1c). Here, the prime denotes the regression coefficient against the DJF Niño3.4 SST index, representing anomalies in tropospheric specific humidity, air temperature, geopotential height and zonal wind shown afterward. The MME $q'_{sfc}$ increased by 33 ± 19% from $P_{ctrl}$ to $P_{warm}$ (hereafter, the numbers indicate the MME mean ± one standard deviation range), which is equivalent to 8.9 ± 5.1% amplification per 1-K background warming in the tropics. The regression of DJF tropical-mean (10° S–10° N) 200-hPa temperature against the Niño3.4 SST index also increased robustly in every model from $P_{ctrl}$ to $P_{warm}$ (Fig. 1d), by as much as 49.6 ± 32.8% (or 13.4 ± 8.8% per 1-K background tropical warming), highlighting the dramatic strengthening of the upper

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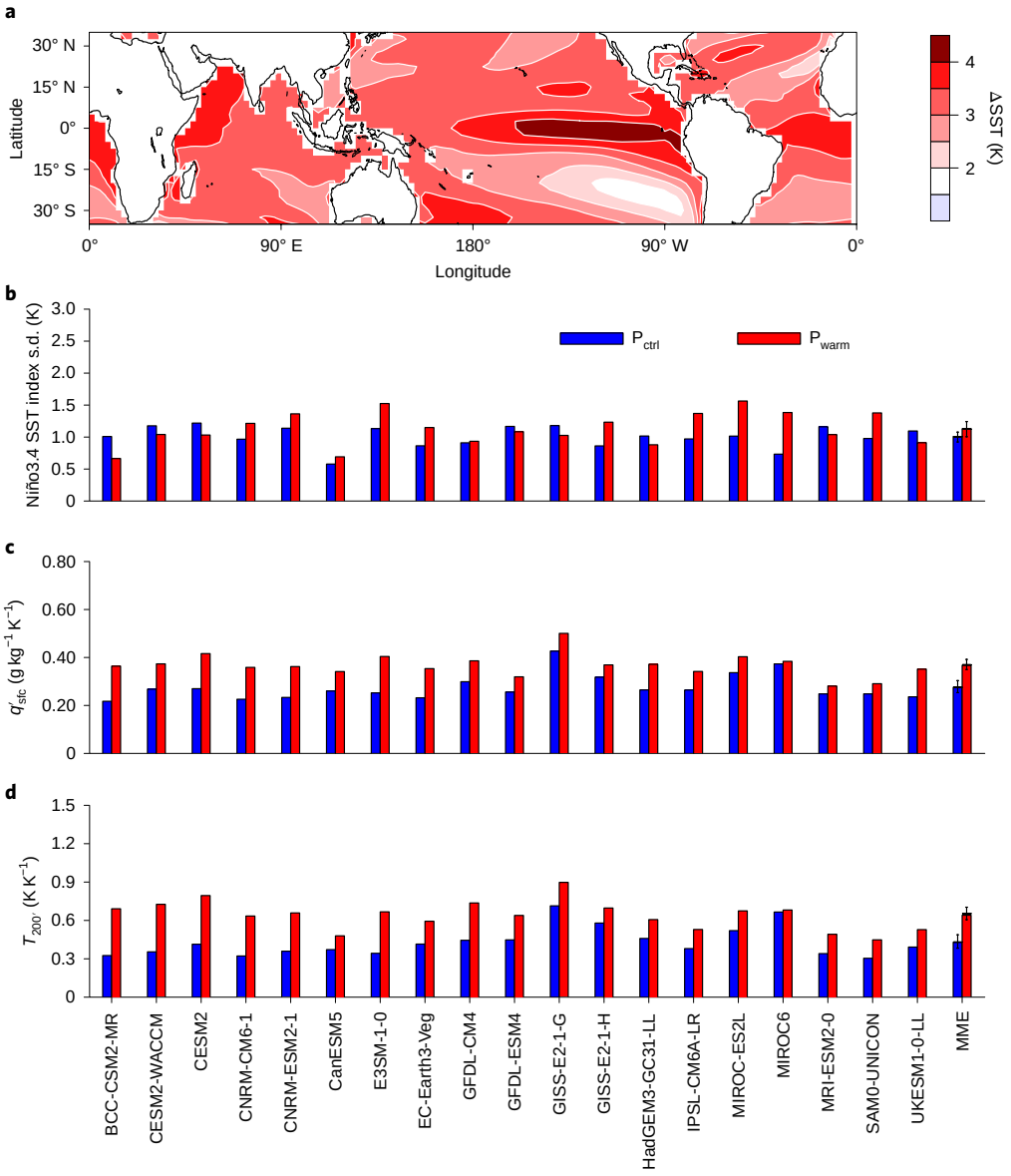



**Fig. 1 | Projected changes of ENSO humidity and air temperature. a**, Changes in MME mean boreal winter (DJF) climatological SST from $P_{ctrl}$ to $P_{warm}$. **b**, Comparison of the standard deviation of the DJF Niño3.4 SST index over $P_{ctrl}$ (blue) and $P_{warm}$ (red) in each model and the MME. **c**,**d**, Comparison of DJF tropical-mean (10° S–10° N) ENSO-driven surface specific humidity ($q'_{sfc}$; **c**) and 200-hPa air temperature ($T_{200}'$; **d**) between $P_{ctrl}$ (blue) and $P_{warm}$ (red). Error bars in the MME mean correspond to the 95% confidence interval (see Methods).

tropospheric temperature response to the ENSO under greenhouse warming. The amplification of ENSO-driven variability in surface specific humidity and upper tropospheric temperature in the tropics was also robust across models in other scenario simulations for both CMIP5 and CMIP6 (Extended Data Figs. 2 and 3).

Why does ENSO variability amplify in tropical air humidity and tropospheric temperature under global warming? We start from the Clausius–Clapeyron expression for the saturation vapour pressure: $dq_s/dT = q_s/L(R_vT^2)$, where $L$ is the latent heat of vaporization, $R_v$ is the gas constant, $q_s$ is the saturation vapour pressure and $T$ is the air temperature. The relationship between $q_s$ and $T$ is nearly exponential and the value of $dq_s/dT$ is approximately proportional to $q_s$[37,38]. Under global warming, $q_s$ increases by ~7% for each 1-K increase in temperature[39], as does $dq_s/dT$. Since the tropical atmosphere tends to maintain a fixed tropospheric relative humidity as it warms[39], the increase of $dq_s/dT$ means a strengthening response of vapour to air temperature variability. Thus, we can deduce that the ENSO drives a larger tropical lower tropospheric moisture anomaly in a warmer world, even if the SST variability remains unchanged.

CGCM projections agree well with the theoretical deduction. Figure 2a shows the percentage changes of DJF tropical-mean (10° S–10° N) surface $dq_s/dT$ against the DJF tropical-mean surface air temperature increase ($\Delta\bar{T}$) in each model and the MME mean. The value of $dq_s/dT$ was calculated as the regression of surface saturation vapour pressure against surface air temperature in each of the four periods (400–449 model years in the pre-industrial control runs and 0–49, 50–99 and 100–149 model years in the 1% yr$^{-1}$ [$CO_2$] runs), which represent the reference and the three different warming stages. The linear fit has a slope of ~6.2% K$^{-1}$, which agrees well with the Clausius–Clapeyron scaling. The change in the response of DJF tropical surface specific humidity to air temperature ($dq/dT$) also generally obeys the Clausius–Clapeyron scaling, at a rate of

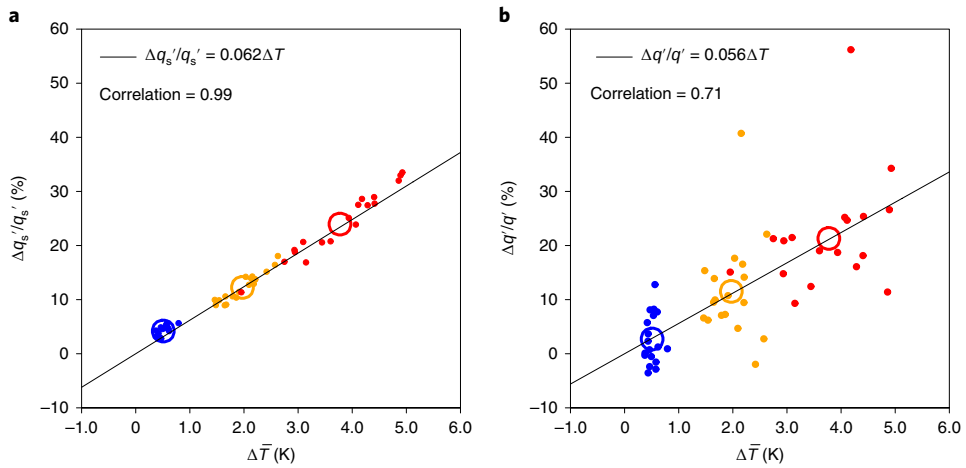



**Fig. 2 | Projected changes in the tropical humidity response to air temperature under global warming. a**, Percentage changes in DJF tropical-mean $dq_s/dT$ (abbreviated as $q_s'$) against the increase of tropical climatological surface temperature ($\Delta\bar{T}$) in each model (dots) and the MME (open circles) during three periods. The blue, orange and red colours denote the changes in the periods 0–49, 50–99 and 100–149 model years, respectively, in the 1% $yr^{-1}$ [$CO_2$] runs from $P_{ctrl}$. The value of $dq_s/dT$ was calculated as the regression of surface saturation vapour pressure onto surface air temperature in each period. **b**, As in **a**, but for surface-specific humidity.

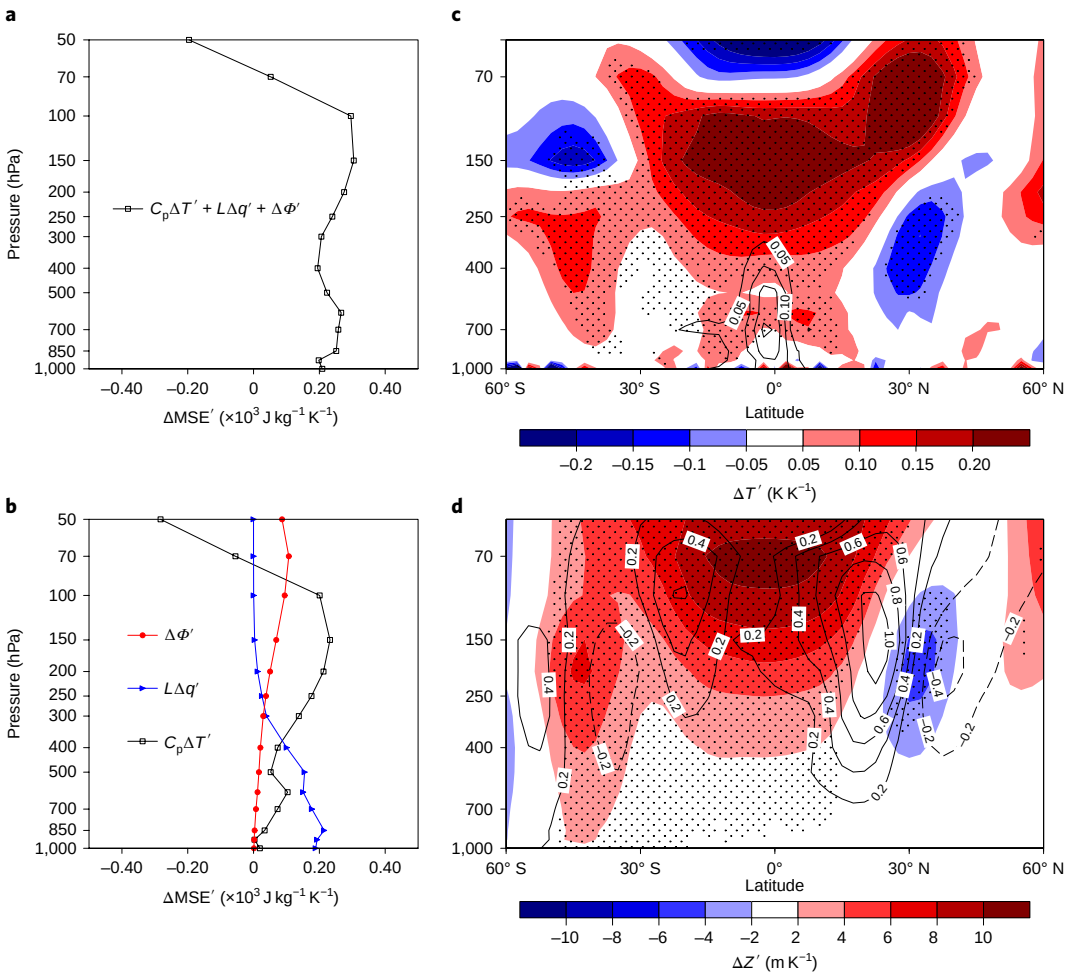



**Fig. 3 | Vertical structure of the changes in ENSO humidity, temperature and circulation. a**, Changes in DJF tropical-mean ENSO-driven moist static energy anomalies (ΔMSE′) in the MME. **b**, Components of the changes in moist static energy anomalies ($C_p\Delta T'$, $L\Delta q'$ and $\Delta\phi'$). **c**, Changes in DJF zonal-mean ENSO-driven specific humidity anomalies ($\Delta q'$; contours; at intervals of 0.05 g $kg^{-1}$ $K^{-1}$) and air temperature anomalies ($\Delta T'$; colours). **d**, Changes in DJF zonal-mean ENSO-driven zonal wind anomalies ($\Delta U'$; black contours; at intervals of 0.2 m $s^{-1}$ $K^{-1}$) and geopotential height anomalies ($\Delta Z'$; colours). In **c** and **d**, stippling indicates that more than 85% of models agree on the sign of the MME.

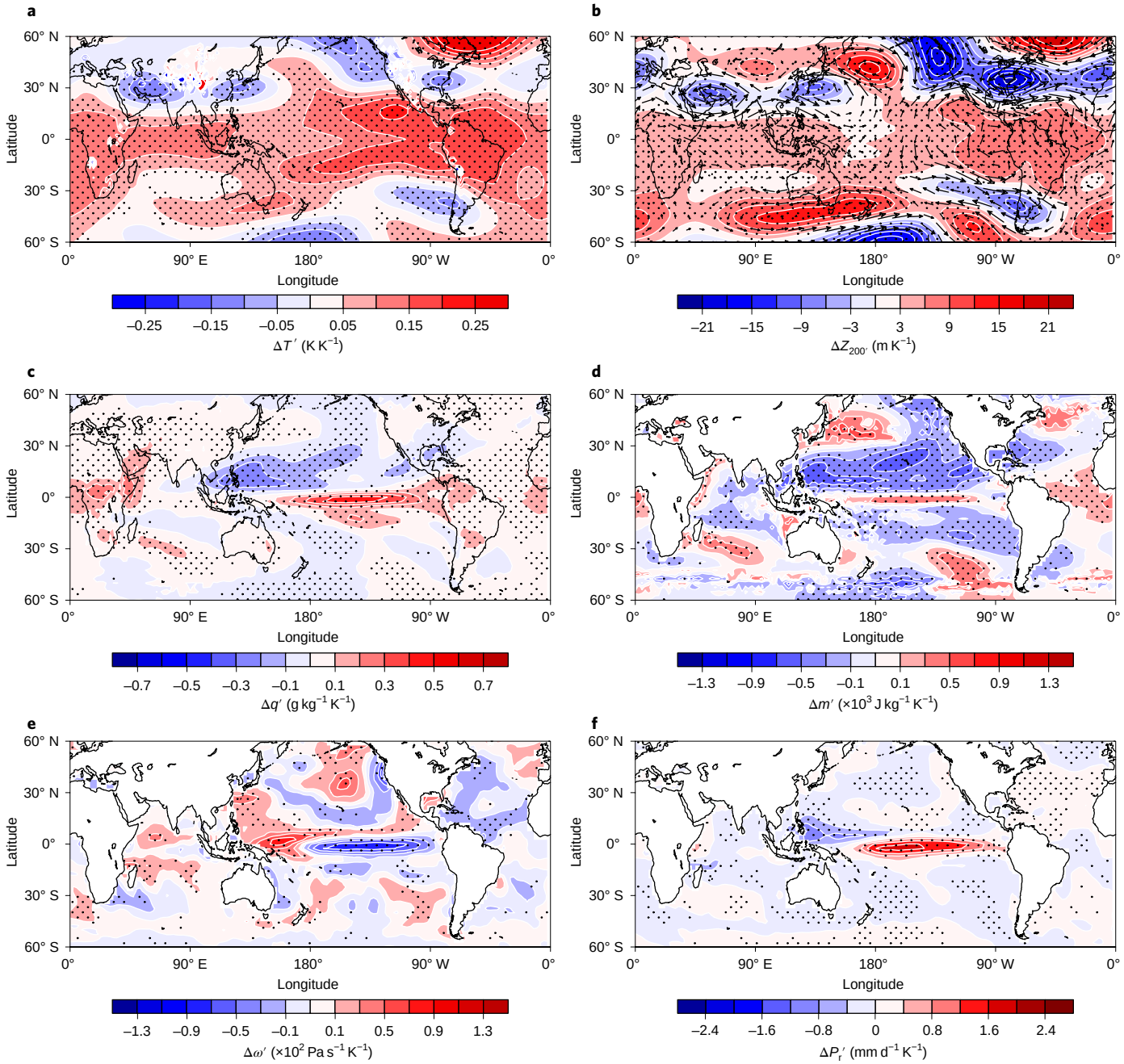



**Fig. 4 | Mechanism for the changes in ENSO atmospheric circulation and precipitation. a**, Changes in the ENSO-driven DJF vertically averaged (850–200 hPa) air temperature anomaly ($\Delta T'$). **b**, Changes in ENSO-driven DJF 200-hPa geopotential height ($\Delta Z_{200}'$; colours) and wind ($\Delta UV_{200}'$; vectors; arrows represent 3 m s$^{-1}$ K$^{-1}$) anomalies. *UV*, zonal (*U*) and meridional (*V*) components of wind. **c**, Changes in ENSO-driven DJF low-level averaged (1,000–500 hPa) specific humidity anomalies ($\Delta q'$). **d**, Changes in ENSO-related DJF moist instability anomalies ($\Delta m'$). **e**, Changes in ENSO-driven DJF 500-hPa vertical pressure velocity anomalies ($\Delta\omega'$). **f**, Changes in ENSO-driven DJF rainfall anomalies ($\Delta P_r'$). Stippling indicates that more than 85% of models agree on the sign of the MME.

~5.6% per 1-K background warming (Fig. 2b). Compared with $dq_s/dT$, the relationship between $dq/dT$ and $\Delta\bar{T}$ has some diversity, probably caused by uncertainty in the relative humidity change. Figure 3a shows the vertical profile of the change in El Niño-induced tropical-mean (10° S–10° N) moist static energy (MSE)[40] (see Methods) anomalies from $P_{ctrl}$ to $P_{warm}$ in the MME. In the troposphere, the El Niño-induced MSE anomaly increases almost uniformly in the vertical profile under global warming, due to moist adiabatic adjustment in convection[11,39]. Unlike MSE itself, the three components (see Methods) in the troposphere do not distribute uniformly in the vertical profile (Fig. 3b). The ENSO-induced humidity anomaly amplifies under global warming, obeying the Clausius–Clapeyron scaling, which is confined to the lower troposphere. To maintain moist adiabats in the tropics, ENSO-induced upper tropospheric temperature anomalies increase proportionally (Fig. 3b). Thus, the Clausius–Clapeyron-based mechanism can explain why ENSO-driven variability will amplify in tropical air humidity and at tropospheric temperature under global warming.

## Atmospheric circulation variability

The enhanced response of tropical air temperature and humidity to the ENSO can result in a major reorganization of the atmospheric circulation. As shown in Fig. 3c, the change of zonal-mean specific humidity anomalies from $P_{ctrl}$ to $P_{warm}$ is mainly confined in the lower troposphere below 500 hPa, while the change of air temperature anomalies peaks at around 150 hPa. By hydrostatic balance, the El Niño-induced geopotential height anomaly strengthens in the upper tropical troposphere, with intensified westerly wind anomalies in the upper troposphere around 20° S and 20° N (Fig. 3d). Previous studies[12,15] showed that subtropical jets in both hemispheres strengthen on their equatorward flanks in El Niño events (Extended Data Fig. 4) and our results indicate that this response will amplify under global warming. We found that the El Niño-induced zonal-mean westerly wind anomalies in the upper troposphere around 20° N increase by ~61.7 ± 60% from $P_{ctrl}$ to $P_{warm}$, or by 16.6 ± 16.2% for each 1-K background warming in the tropics.

Because of equatorial wave adjustments[13,37], the increase of ENSO-induced air temperature anomalies in the free troposphere spread horizontally in the entire tropics (Fig. 4a). Consequently, there are positive changes of El Niño-induced 200-hPa height anomalies within the tropics between 20° S and 20° N, and enhanced westerly wind anomalies along 20° S and 20° N due to the equatorward pressure gradient (Fig. 4b). Poleward of the tropical warming, there is a latitude belt of marked tropospheric cooling in the Northern Hemisphere, probably caused by the changes in the eddy-driven meridional circulation[12]. The cooling is prominent over the Asian continent and North America, which corresponds to robust 200-hPa low-pressure anomalies in these regions. Thus, the pressure gradient south of these regions is stronger than in other regions, and so are the westerly wind anomalies. Compared with the ENSO-induced air temperature and atmospheric circulation anomalies in the present real climate (Extended Data Fig. 5), the anomalies intensify in a warmer climate. Particularly, the westerly wind anomalies over subtropical North America (15° N–30° N, 130° W–30° W) increase from 1.9 ± 0.7 m s$^{-1}$ K$^{-1}$ for $P_{ctrl}$ to 3.3 ± 0.5 m s$^{-1}$ K$^{-1}$ for $P_{warm}$, corresponding to a 19.2 ± 11.3% increase for each 1-K background tropical warming. This intensified circulation response makes the US west coast more susceptible to ENSO varibility[30–33].

## Precipitation variability

The changes in ENSO-driven humidity anomalies are localized. In El Niño, SST and low-level tropospheric air temperature anomalies are positive in the tropical central-eastern Pacific but negative in the tropical Northwest Pacific. Hence, the background greenhouse warming will strengthen both the positive humidity anomalies over the central-eastern Pacific and negative humidity anomalies over the Northwest Pacific (Fig. 4c and Extended Data Fig. 6). The changes of ENSO-related gross moist instability[40] (see Methods) are primarily determined by the changes of ENSO-driven humidity anomalies, with increasing trends over the tropical central-eastern Pacific but decreasing trends in the Northwest Pacific (Fig. 4d). In the tropics, the increase of gross moist instability tends to lead to low-level convergence, and vice versa[40]. Indeed, the changes of ENSO-induced 500-hPa vertical velocity anomalies are spatially consistent with the change of gross moist instability in the tropics, with upwelling change over the equatorial central-eastern Pacific but downwelling change over the Northwest Pacific (Fig. 4e). Therefore, the enhanced response of humidity to air temperature can intensify both the El Niño-driven precipitation decrease over the Northwest Pacific and the increase over the central-eastern equatorial Pacific (Fig. 4f and Extended Data Fig. 6). This explanation does not require spatial variations in mean warming[18,27,29] or ENSO SST anomalies[41]. The relative contributions from the SST pattern effect need to be evaluated in future studies.

Year-to-year ENSO variability is controlled by complicated air–sea feedbacks and affected by the noise arising from internal variability. It proves difficult to determine how ENSO behaviours change under global warming[16,42–46]. Here, we find that the nonlinear response of saturation vapour pressure to temperature change (that is, the Clausius–Clapeyron relationship) causes robust changes in key characteristics of the ENSO under global warming. Specifically, the boreal winter response of tropical humidity, tropospheric upper temperature, the subtropical jet in both hemispheres and tropical Pacific rainfall to the ENSO intensify robustly across the models. As extreme weather often results from ENSO-induced anomalous atmospheric circulation and temperature[47–49], the intensification of ENSO-driven atmospheric variability offers a key step to project how ENSO-related extreme weather will change in the future, which deserves further studies.

This study shows that the Clausius–Clapeyron-based mechanism is an important constraint for ENSO projections. This mechanism is based on the change in the sensitivity of vapour response to temperature but not on the change in the mean-state vapour, as has been widely used in previous studies[39,50]. The response of tropospheric humidity to SST variability also intensifies elsewhere (for example, the tropical Indian Ocean and Atlantic; Extended Data Figs. 7 and 8) under greenhouse warming, suggesting that our method can be applied to projecting changes in other tropical modes.

## Online content

Any methods, additional references, Nature Research reporting summaries, source data, extended data, supplementary information, acknowledgements, peer review information; details of author contributions and competing interests; and statements of data and code availability are available at https://doi.org/10.1038/s41561-021-00730-3.

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## Methods

**Datasets and index definition.** We used the pre-industrial control runs and 1% $yr^{-1}$ [$CO_2$] increase runs from 19 CGCMs of CMIP6. These were BCC-CSM2-MR, CESM2-WACCM, CESM2, CNRM-CM6-1, CNRM-ESM2-1, CanESM5, E3SM-1-0, EC-Earth3-Veg, GFDL-CM4, GFDL-ESM4, GISS-E2-1-G, GISS-E2-1-H, HadGEM3-GC31-LL, IPSL-CM6A-LR, MIROC-ES2L, MIROC6, MRI-ESM2-0, SAM0-UNICON and UKESM1-0-LL. To confirm the results, we performed two additional analyses. One was to analyse the changes from the historical scenario run (1950–1999) to the SSP585 run (2050–2099) in 30 CMIP6 models (Extended Data Fig. 3). The other was to analyse the changes from the historical scenario run (1950–1999) and the representative concentration pathway 8.5 run (2050–2099) in 18 CMIP5 models (Extended Data Fig. 2). The 30 CMIP6 models were ACCESS-CM2, ACCESS-ESM1-5, BCC-CSM2-MR, CESM2-WACCM, CESM2, CNRM-CM6-1-HR, CNRM-CM6-1, CNRM-ESM2-1, CanESM5-CanOE, GFDL-CM4, CanESM5, EC-Earth3-Veg, FGOALS-f3-L, FGOALS-g3, GFDL-CM4, GFDL-ESM4, GISS-E2-1-G, INM-CM4-8, INM-CM5-0, IPSL-CM6A-LR, KACE-1-0-G, MCM-UA-1-0, MIROC-ES2L, MIROC6, MPI-ESM1-2-HR, MPI-ESM1-2-LR, MRI-ESM2-0, NorESM2-LM, NorESM2-MM and UKESM1-0-LL. The 18 CMIP5 models were ACCESS1-0, BNU-ESM, CCSM4, CESM1-BGC, CNRM-CM5, CSIRO-Mk3-6-0, CanESM2, IPSL-CM5A-LR, IPSL-CM5A-MR, IPSL-CM5B-LR, MIROC-ESM-CHEM, MIROC-ESM, MIROC5, MRI-CGCM3, NorESM1-ME, NorESM1-M, bcc-csm1-1-m and bcc-csm1-1. More detailed online model documentations for the CMIP6 and CMIP5 models are available at https://pcmdi.llnl.gov/CMIP6/ and https://esgf-node.llnl.gov/search/cmip5/, respectively.

The MME is defined as the simple average of the 19 models. SST, precipitation, air temperature, geopotential height, winds, air specific humidity and vertical pressure velocity at 500 hPa were used. We constructed a DJF Niño3.4 SST index by averaging DJF SST anomalies in the domain of 5° S–5° N and 90°–150° W to indicate ENSO variations. ENSO-driven anomalies of specific humidity, air temperature, geopotential height, vertical pressure velocity and winds were calculated by regression on the DJF Niño3.4 SST index so that they represent variable anomalies per 1 K Niño3.4 SST index change. Before the regression analyses, the linear trend was removed. We also used a scaling method[36] to detrend the data and the result was almost the same as when applying a linear detrending method, indicating that the result is not sensitive to method choices (Extended Data Fig. 1).

For comparison, we calculated the observed ENSO-driven atmospheric anomalies based on the 2.5° latitude × 2.5° longitude daily National Centers for Environmental Prediction/Department of Energy version 2 Reanalysis[51] and Global Precipitation Climatology Project monthly precipitation datasets[52] and the 1° latitude × 1° longitude monthly Hadley Center SST[53] over 1979–2018. The results are shown in the Extended Data Figs. 4–6.

**Moist static energy and moist instability analyses.** MSE[40] is defined as $\mathrm{MSE} = C_p T + \phi + Lq$, where $L$ is the latent heat of condensation, $C_p$ is the specific heat capacity, $T$ is the air temperature, $\phi$ is the geopotential and $q$ is the air specific humidity. Thus, the change of ENSO-driven MSE anomalies ($\Delta$MSE′) can be represented by $C_p\Delta T' + L\Delta q' + \Delta\phi'$. Here, $\Delta T'$, $\Delta q'$ and $\Delta\phi'$ denote the change of ENSO-driven anomalies of air temperature, specific humidity and geopotential, respectively. Since large-scale divergence tends to have a simple vertical structure in the tropics, with one sign in the lower troposphere and the opposite sign in the upper troposphere, for simplicity, we define gross moist instability ($m$) as the difference in moist static energy between the lower troposphere (1,000–500 hPa) and upper troposphere (500–200 hPa) following previous studies[13,37,40].

**Statistical significance.** We applied a 10,000-resampling bootstrap method to test whether the variance changes of the DJF Niño3.4 SST index, ENSO-related tropical-mean surface humidity and 200 hPa air temperature were statistically significant. For each of the three variables, its values in the 19 models were resampled randomly to construct 10,000 realizations of mean values. In this random resampling process, any model was allowed to be selected again. The 10,000 mean values were sorted in ascending order and the 250th and 9,750th values were set as the 2.5% lower confidence bound and the 97.5% upper confidence bound. Between the two bounds was the 95% confidence interval. In the observational analyses, the confidence level was based on a two-tailed Student's $t$-test.

## Data availability

The CMIP6 data are available at https://pcmdi.llnl.gov/CMIP6/. The CMIP5 data are available at https://esgf-node.llnl.gov/search/cmip5/. The version 2 National Centers for Environmental Prediction/Department of Energy reanalysis and Global Precipitation Climatology Project precipitation data are from http://www.esrl.noaa.gov/psd/data/gridded/. Hadley Center SST data were provided by the Met Office Hadley Center (https://www.metoffice.gov.uk/hadobs/hadisst/).

## Code availability

The code associated with this paper is available on request from K.H.

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## Author contributions

K.H. and G.H. conceived the study, performed the analyses, built the mechanism and wrote the paper. P.H., Y.K. and S.-P.X. contributed to improving the paper and assisted in interpretation of the results.

## Competing interests

The authors declare no competing interests.

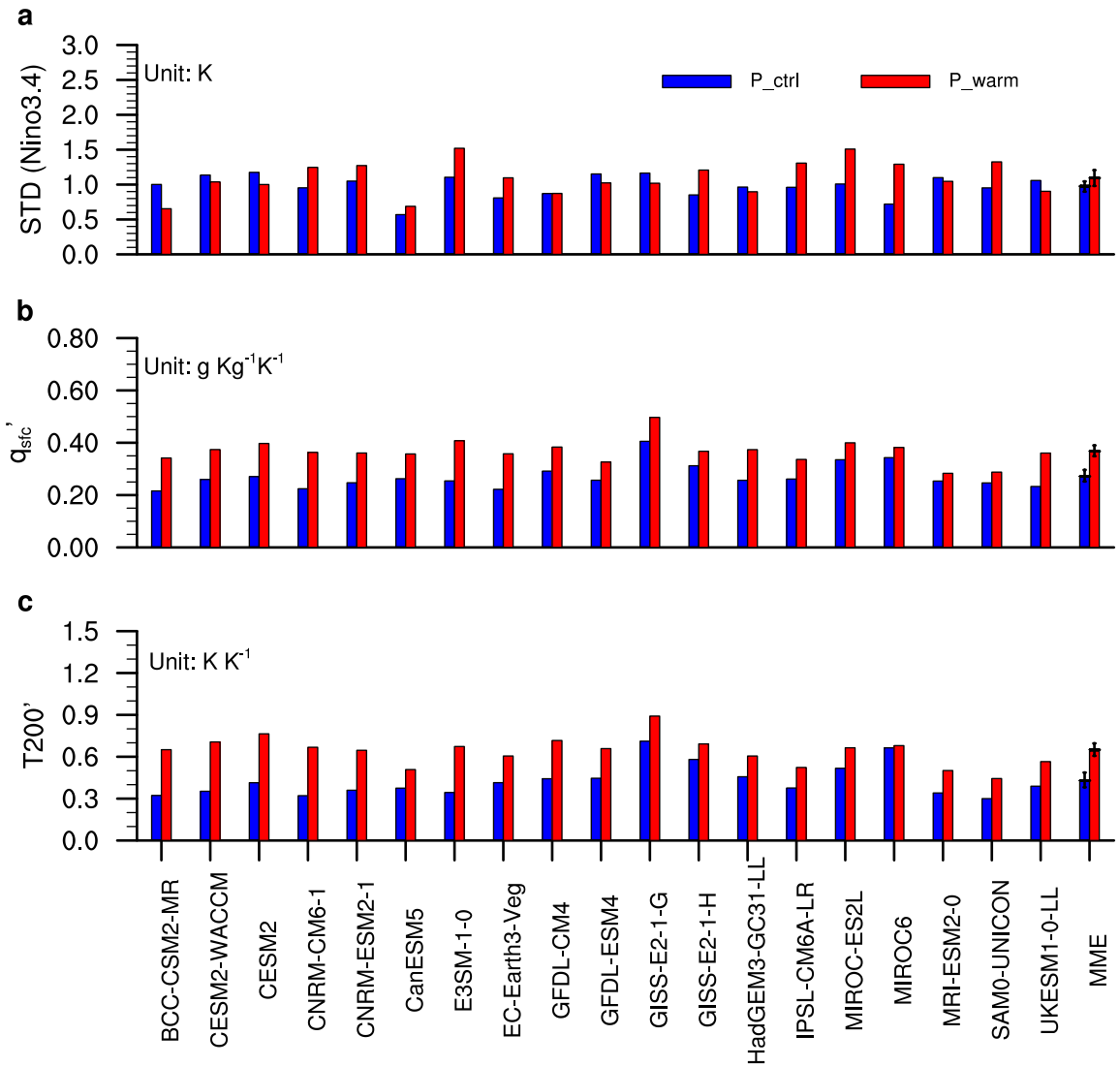
## Additional information

**Extended data** is available for this paper at https://doi.org/10.1038/s41561-021-00730-3.

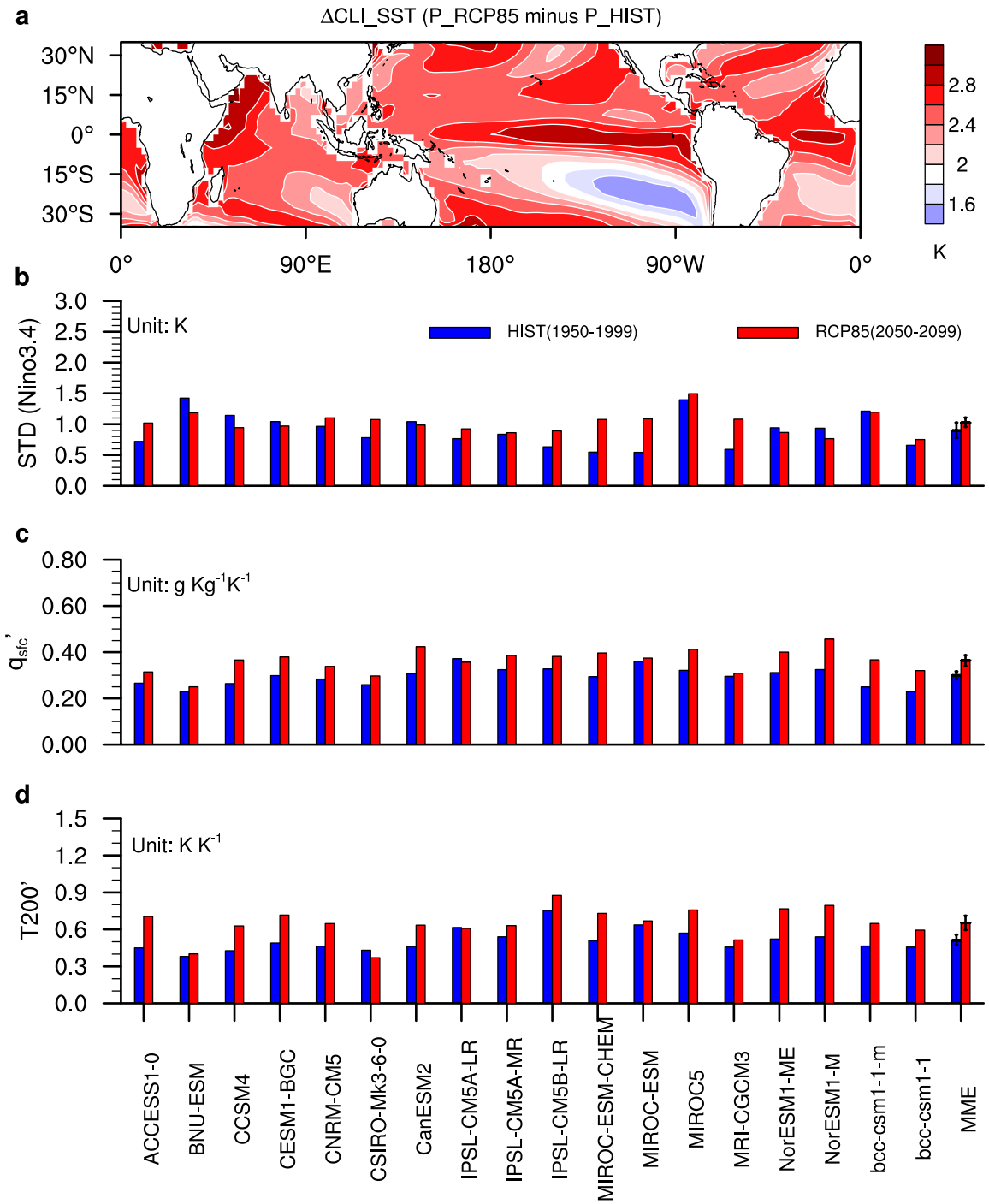
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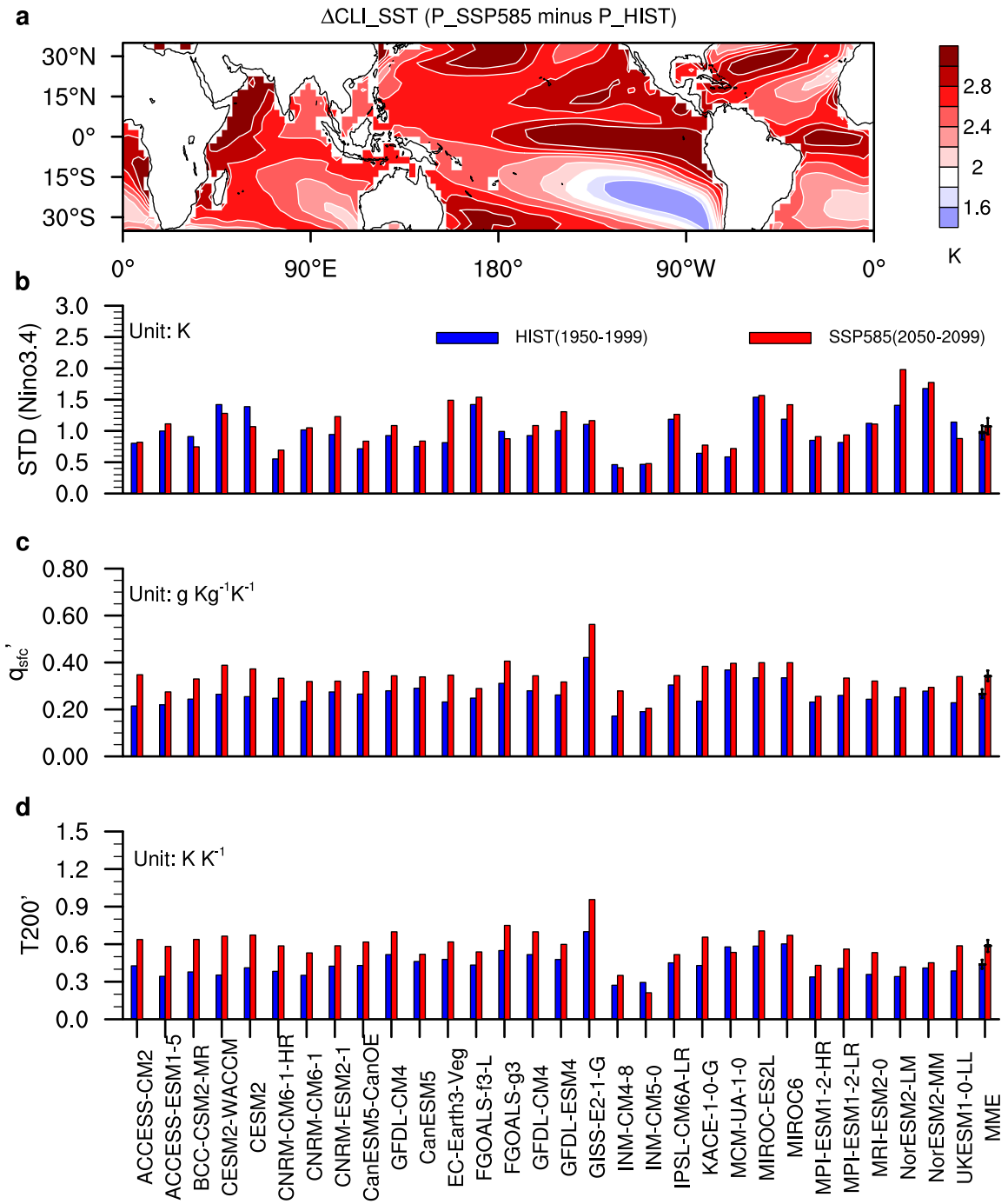
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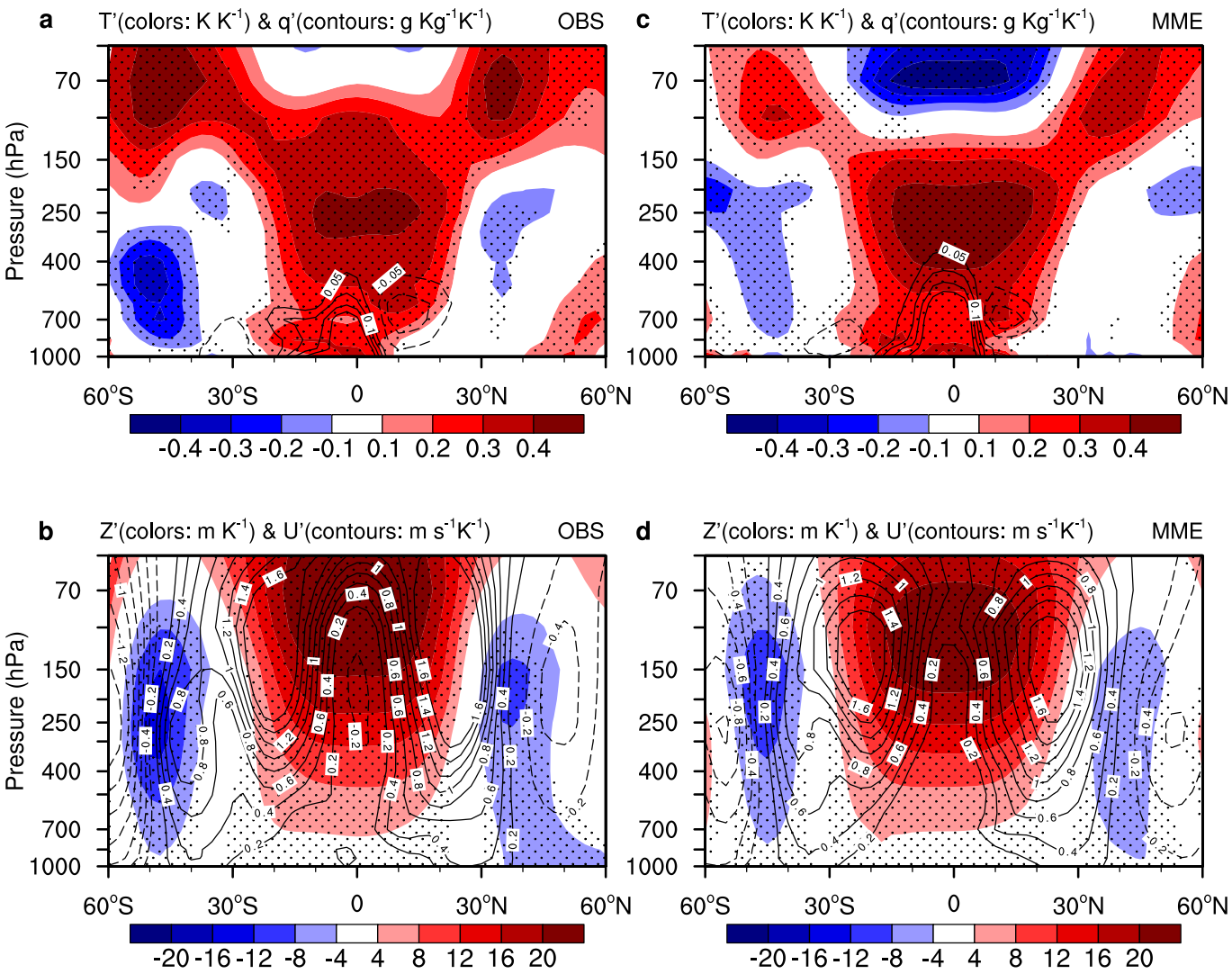

**Extended Data Fig. 1 | Projected change in ENSO-driven variability in boreal winter tropical SST, humidity and air temperature.** The analysis is the same as that in Fig. 1 except for applying a "scaling" method (see Method) for detrending data prior to regressions. The result is almost the same as that in Fig. 1, indicating the result is not sensitive to the selection of method for detrending.



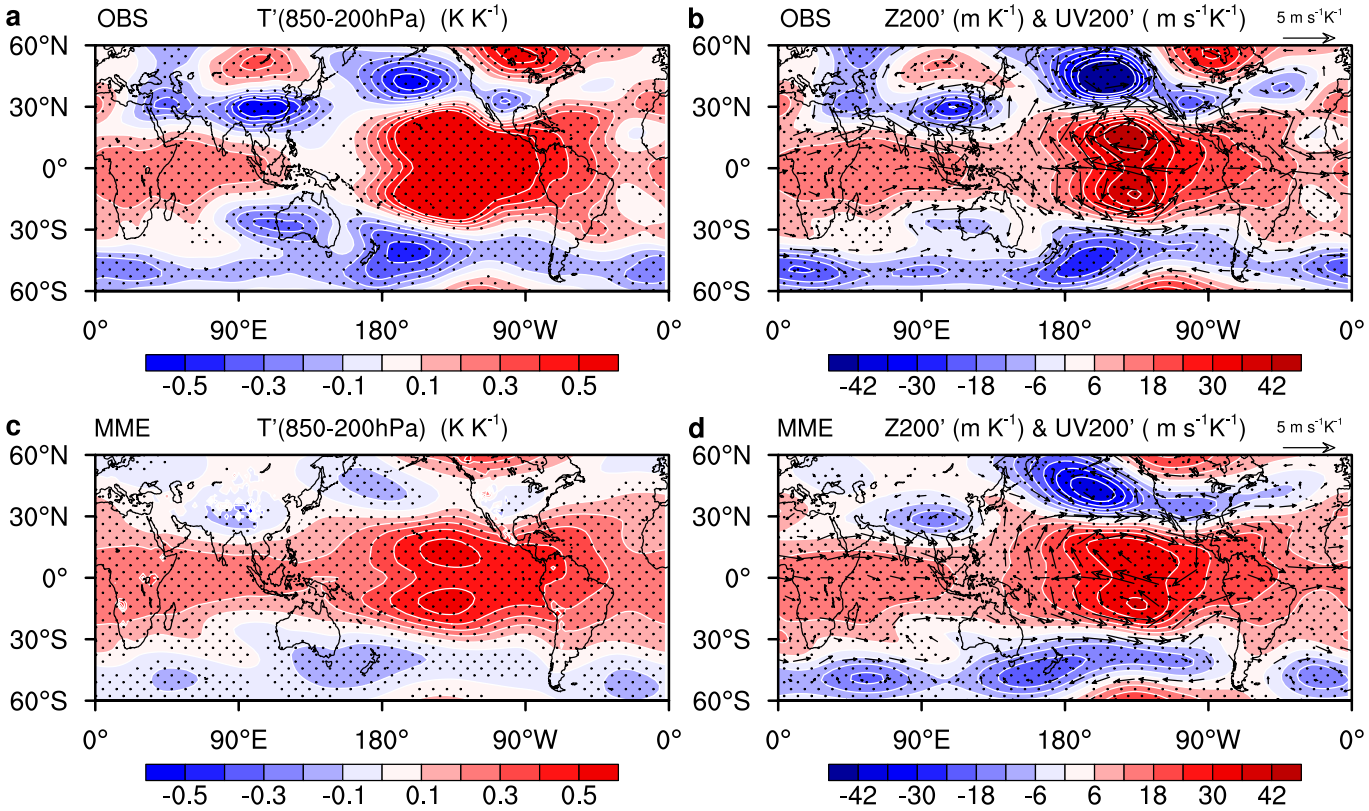

**Extended Data Fig. 2 | Projected change in ENSO SST, humidity and air temperature in 18 CMIP5 models (see Methods) from 1950-1999 in historical experiments to 2050-2099 in RCP8.5 experiments. a**, Changes in DJF climatological SST in the MME. **b**, Comparison of the standard deviation of DJF Niño 3.4 SST index over the two periods in each model and the MME. **c, d**, Comparison of DJF tropical mean (10°S-10°N) ENSO-driven surface specific humidity ($q'_{sfc}$; **c**) and 200-hPa air temperature ($T'_{200}$; **d**) between the two periods. Error bars in the MME mean correspond to the 95% confidence interval (see Methods). The MME $q'_{sfc}$ and $T'_{200}$ increases by 9.6±6.1% and 11.5±8.4% respectively for per 1K background warming in the tropics.



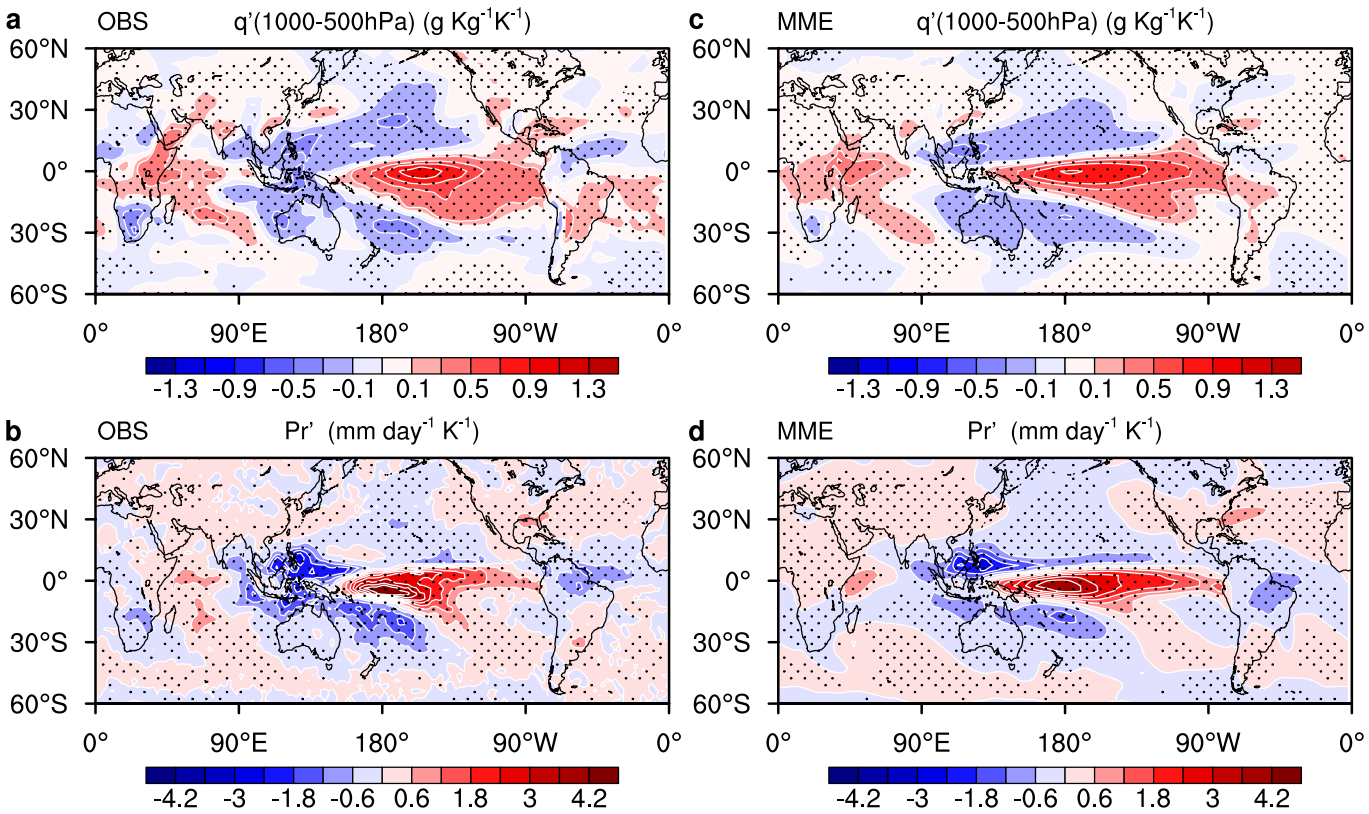

**Extended Data Fig. 3 | Projected change in ENSO SST, humidity and air temperature in 30 CMIP6 models (see Methods) from 1950-1999 in historical experiments to 2050-2099 in SSP585 experiments. a**, Changes in DJF climatological SST in the MME. **b**, Comparison of the standard deviation of DJF Niño 3.4 SST index over the two periods in each model and the MME. **c, d**, Comparison of DJF tropical mean (10°S-10°N) ENSO-driven surface specific humidity ($q'_{sfc}$; **c**) and 200-hPa air temperature ($T'_{200}$; **d**) between the two periods. Error bars in the MME mean correspond to the 95% confidence interval (see Methods). The MME $q'_{sfc}$ and $T'_{200}$ increases by 10.2±6.0% and 12.2±8.5% respectively for per 1K background warming in the tropics.



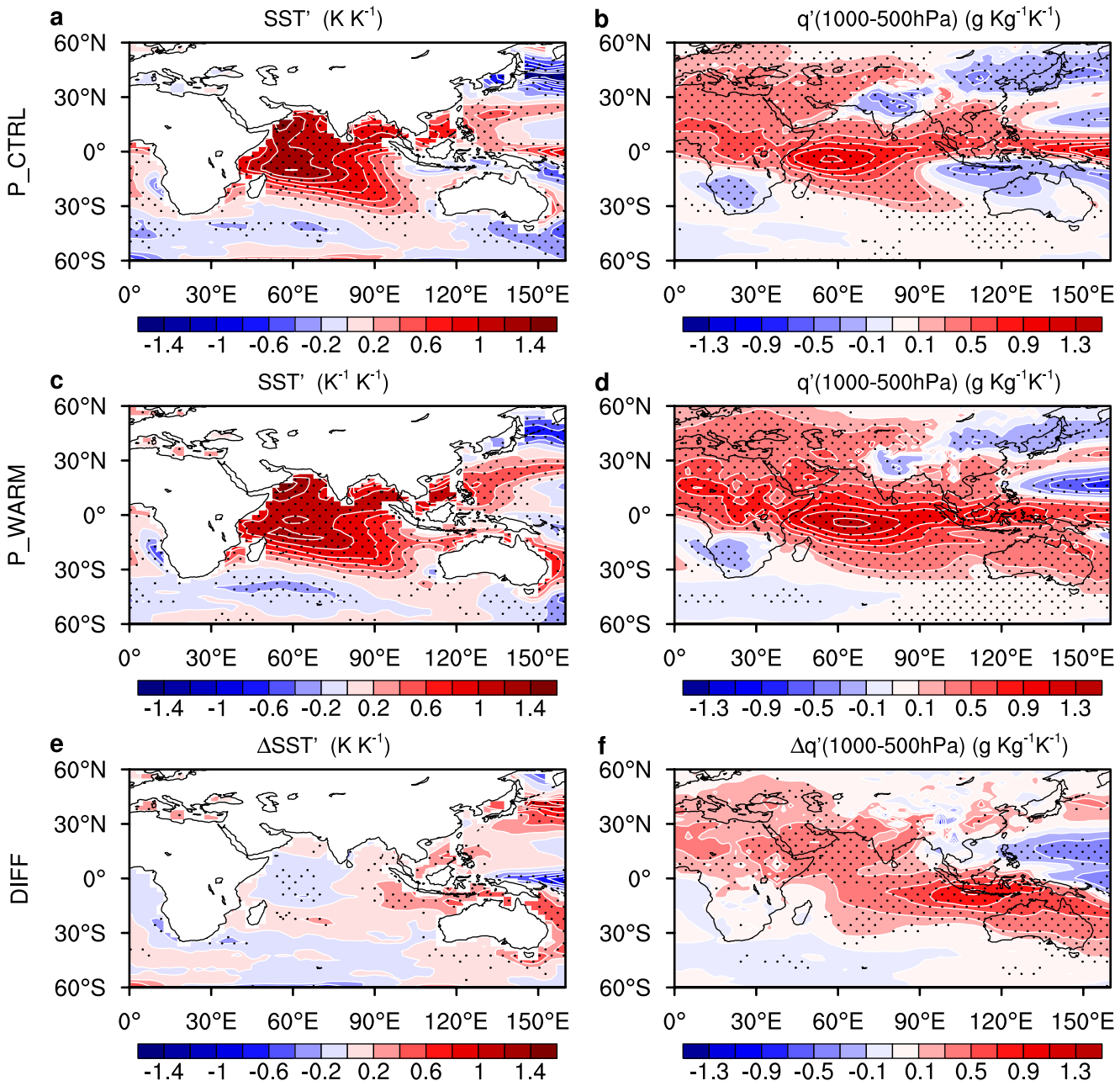

**Extended Data Fig. 4 | Vertical structure of ENSO-driven anomalies of boreal winter air humidity, temperature and circulation. a**, DJF zonally mean ENSO-driven specific humidity anomalies (q′; contours; at interval of 0.05 g kg$^{-1}$ K$^{-1}$) and air temperature anomalies (T′; colors) during 1979-2018 in the observations. **b**, DJF zonally mean ENSO-driven zonal wind anomalies (U′; black contours; at interval of 0.2 m s$^{-1}$ K$^{-1}$) and geopotential height anomalies (Z′; colors) during 1979-2018 in the observations. **c**-**d**, similar to **a** and **b** but for the MME during 400-449 model years in the PI control runs. Stippling in **a** and **b** denotes passing 95% confidence level based on a two-tailed Student's t test, and in **c** and **d** indicates that more than 85% of models agree on the sign of the MME.



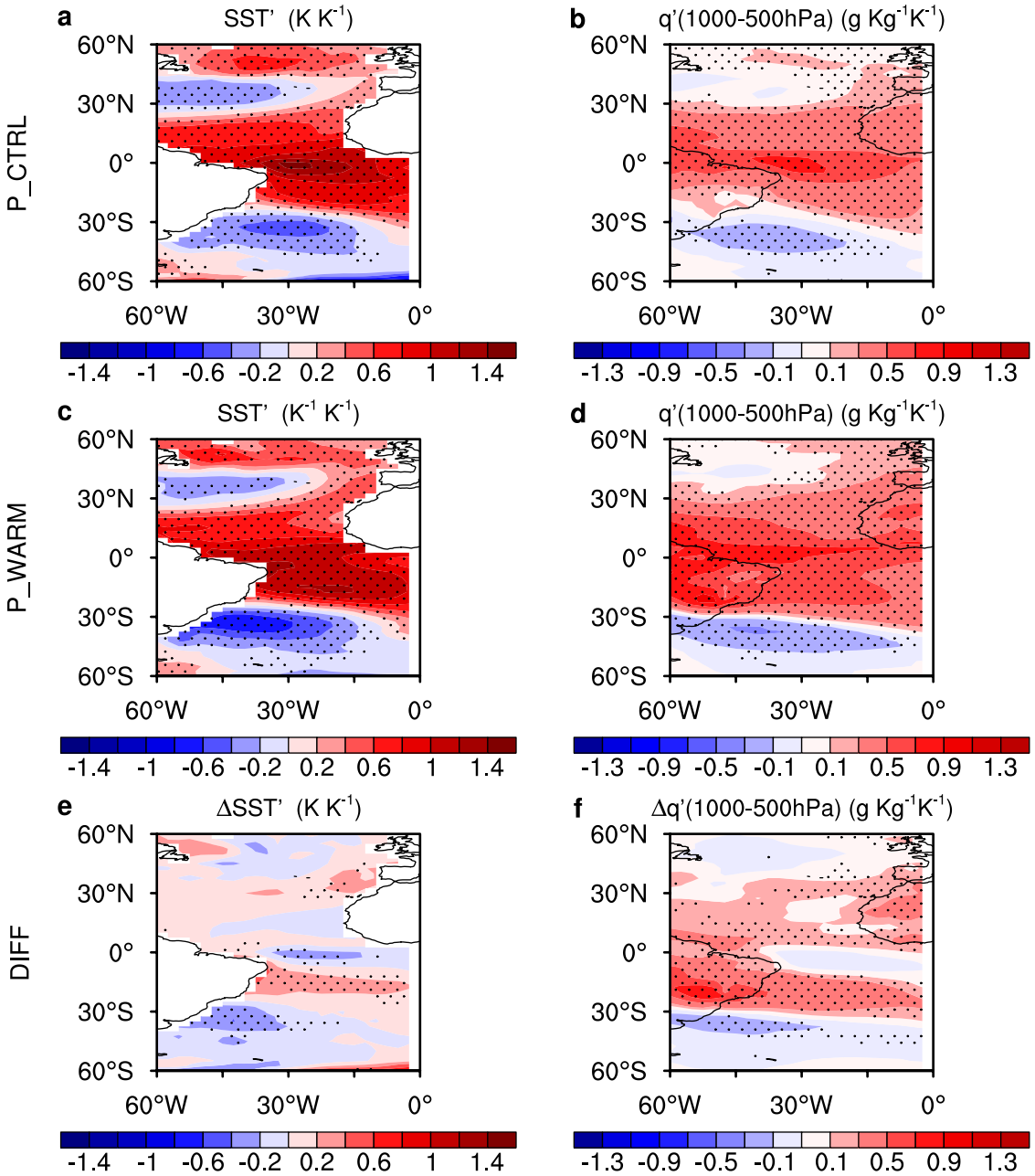

**Extended Data Fig. 5 | ENSO-driven anomalies of boreal winter tropospheric air temperature, 200hPa geopotential height and circulation. a** and **c**, Vertically averaged (850-200 hPa) ENSO-driven DJF air temperature anomalies (T′). **b** and **d**, 200-hPa ENSO-driven DJF geopotential height (Z200′; colors) and wind (UV200′; vectors) anomalies. Anomalies in **a** and **b** are derived from the observations during 1979-2018, while in **c** and **d** are from the MME during 400-449 model years in the PI control runs. Stippling in **a** and **b** denotes passing 95% confidence level based on a two-tailed Student's t test, and in **c** and **d** indicates that more than 85% of models agree on the sign of the MME.




**Extended Data Fig. 6 | ENSO-driven DJF anomalies in boreal winter specific humidity and rainfall. a** and **c**, ENSO-driven DJF low-level 1000-500hPa vertically averaged specific humidity anomalies (q'). **b** and **d**, ENSO-driven DJF rainfall anomalies (Pr'). Anomalies in **a** and **b** are derived from the observations during 1979-2018, while in **c** and **d** are from the MME during 400-449 model years in the PI control runs. Stippling in **a** and **b** denotes passing 95% confidence level based on a two-tailed Student's t test, and in **c** and **d** indicates that more than 85% of models agree on the sign of the MME.




**Extended Data Fig. 7 | Changes in the response of specific humidity to tropical Indian Ocean SST variability under global warming. a** and **b**, The regression of June-to-August (JJA) mean SST **(a)** and 1000-500hPa vertically averaged specific humidity **(b)** onto a tropical Indian Ocean SST index in the period of P_ctrl. The tropical Indian Ocean SST index is defined as averaging JJA SST anomalies in the domain of 10°S-10°N, 40°E-100°E. **c** and **d**, Similar to **a** and **b** but for the period of P_warm. **e** and **f**, Differences between the two periods (P_warm- P_ctrl). Stippling indicates that more than 85% of models agree on the sign of the MME. The result shows that the response of specific humidity to tropical Indian Ocean SST variability strengthens under global warming.




**Extended Data Fig. 8 | Changes in the response of specific humidity to tropical Atlantic SST variability under global warming. a** and **b**, The regression of JJA SST (**a**) and 1000-500hPa vertically averaged specific humidity (**b**) onto a tropical Atlantic SST index in the period of P_ctrl. The tropical Atlantic SST index is defined as averaging JJA SST anomalies in the domain of 10°S-10°N, 60°W-0°. **c** and **d**, Similar to **a** and **b** but for the period of P_warm. **c** and **d**, Differences between the two periods (P_warm- P_ctrl). Stippling indicates that more than 85% of models agree on the sign of the MME. The result shows that the response of specific humidity to tropical Atlantic SST variability strengthens under global warming.

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# Opposite response of strong and moderate positive Indian Ocean Dipole to global warming

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**A strong positive Indian Ocean Dipole (pIOD) induces weather extremes such as the 2019 Australian bushfires and African floods. The impact is influenced by sea surface temperature (SST), yet models disagree on how pIOD SST may respond to greenhouse warming. Here we find increased SST variability of strong pIOD events, with strong equatorial eastern Indian Ocean cool anomalies, but decreased variability of moderate pIOD events, dominated by western warm anomalies. This opposite response is detected in the Coupled Model Inter-comparison Project (CMIP5 and CMIP6) climate models that simulate the two pIOD regimes. Under greenhouse warming, the lower troposphere warms faster than the surface, limiting Ekman pumping that drives the moderate pIOD warm anomalies; however, faster surface warming in the equatorial western region favours atmospheric convection in the west, strengthening equatorial nonlinear advection that forces the strong pIOD cool anomalies. Climate extremes seen in 2019 are therefore likely to occur more frequently under greenhouse warming.**

A positive Indian Ocean Dipole (pIOD) refers to a sea surface temperature (SST) pattern that features warm anomalies in the equatorial western Indian Ocean and cold anomalies in the east, described by a positive dipole mode index (DMI)[1]. A pIOD event usually peaks in austral spring (September, October, November; SON) but starts to develop in austral winter, with an initial cooling off Sumatra–Java suppressing the local atmospheric convection there and generating anomalous southeasterly winds. The anomalous winds promote upwelling and lift the thermocline that in turn reinforces the initial cooling[1,2], a process referred to as local Bjerknes feedback. During a strong pIOD, the growth of cool anomalies causes a northwestward shift of atmospheric convection and a westward extension of the southeasterly trades that usually flow over the southern tropical Indian Ocean in austral winter and spring. The anomalous easterlies extend across the equatorial Indian Ocean, where weak westerlies normally prevail[3] (Extended Data Fig. 1). The anomalous easterly winds promote warm SST anomalies, convergence, and rainfall in the equatorial western Indian Ocean[1–3].

The altered ocean–atmosphere circulations induce weather extremes in Indian Ocean-rim countries[1–9], such as droughts, heatwaves and bushfires in Indonesia and Australia[4–7], floods and malaria outbreaks in East Africa[9,10], and coral reef death across western Sumatra[11]. During the 1997 strong pIOD event[2] (Extended Data Fig. 1), floods in the African countries led to several thousand deaths and hundreds of thousands of people displaced, whereas wildfires in the eastern Indian Ocean regions affected the lives of tens of millions[3] and the global carbon budget[12]. Similar severe impacts also occurred during the 2019 strong pIOD event[13], and Australia experienced a devastating bushfire season, referred to as the Black Summer, with a burnt area of more than 170,000 km$^2$, 30 people killed and 3,000 buildings destroyed[14].

Determining how pIOD SST may change under greenhouse warming is one of the most important issues in climate science; however, for several climate model generations, there has been either a lack of intermodel consensus or no statistically significant change using the conventional DMI (refs. [15–17]).

## Long-standing uncertainty in projected pIOD SST change

Uncertainty in projected pIOD SST change is despite a strong intermodel consensus on a mean state change in the tropical Indian Ocean, which features equatorial easterly wind trends and a faster warming rate in the northwest than in the southeast of the tropical Indian Ocean, a shallowing equatorial thermocline[15–19], and some changes in characteristics of strong pIOD[20,21]. First, the faster warming in the west equatorial Indian Ocean facilitates an increased frequency of a westward-shifted atmospheric convergence zone from the east and hence there are more events with extreme impacts, even if SST variability does not change[20]. Second, the enhanced north-minus-south SST gradient inhibits the seasonal southward shift of the tropical Indian Ocean Intertropical Convergence Zone[21], leading to an anomalously weak convergence in the eastern Indian Ocean, particularly during a strong pIOD (Extended Data Figs. 1 and 2d–f). Third, although a projected thermocline shallowing in the eastern Indian Ocean increases the role of the relatively cold subsurface water through an increased thermocline response to winds, this is offset by a weakened response of equatorial winds to SST anomalies. The weaker wind response occurs because the lower atmosphere becomes more stable, as the low troposphere warms faster than the surface under greenhouse warming[16].

Palaeoclimate proxies show that pIOD was more intense during the past warmer period of the middle Holocene than in the twentieth century[7], and there is already an increase in the frequency and strength of pIOD events during the second half of

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the twentieth century[8]; however, climate model projections of the response of pIOD SST variability to greenhouse warming remain uncertain[15]. Using the DMI determined from the observed poles assumes that the pIOD possesses little pattern diversity and that the simulated location of pIOD SST anomaly centres across models is the same as the observed. Here we show that there are two pIOD regimes—thus requiring at least two separate indices—and that their associated anomaly patterns differ vastly from one model to another. Furthermore, there is a robust pIOD SST change if we focus on SST variability patterns and centres unique to each individual model.

## Dynamics of the two distinct pIOD regimes in observations

Previous studies have shown that pIOD events display different flavours[20,22–25]. During a strong pIOD, the anomalous cooling off Java–Sumatra expands northward and then westward along the equator[20], inhibiting the seasonal southward shift of the intertropical convergence zone towards the equatorial eastern Indian Ocean and inducing a westward shift of the equatorial eastern Indian Ocean convergence zone[21], generating strong easterlies along the equator (Extended Data Fig. 1d–f). These winds in turn enhance the equatorial west-minus-east SST gradient through a positive nonlinear zonal and vertical feedback, dramatically increasing the anomalous cooling in the equatorial eastern Indian Ocean[20].

During moderate pIOD events such as in 1982, Bjerknes feedback drives moderate cooling off Sumatra–Java (Extended Data Fig. 2a–c); to the west, the dominant forcing of the broad area of warming includes an anomalous Ekman pumping that arises from westward propagation of off-equatorial downwelling Rossby waves that are generated by equatorial anomalous easterlies in the eastern Indian Ocean during the development phase, and their subsequent reflection at the western boundary as downwelling Kelvin waves[26]. Heat budget analysis shows that Ekman pumping is the dominant forcing of the warm anomalies (Extended Data Fig. 3, see also the 'Heat budget analysis' section in the Methods).

## Depiction of the two distinct pIOD regimes

The large difference in the spatial pattern and the dynamics between the strong and moderate pIOD indicates that at least two indices are needed to distinguish them[13,20,27]. To construct the indices, we use the first two principal modes of an empirical orthogonal function (EOF) analysis of SON-averaged SST anomalies that cover the 1982–2015 period, each described by a principal spatial pattern and a principal component (PC) time-series that is scaled to have a variance of unity (see the 'Data, models and EOF analysis' section in the Methods). At their positive phase, EOF1 shows a cold-anomaly centre off Sumatra–Java, which is accompanied by warming in much of the tropical Indian Ocean basin, analogous to that associated with the DMI (correlation of 0.90); this is accompanied by a shoaled equatorial thermocline in the east but a deepened equatorial thermocline to the west (Fig. 1a). EOF2 exhibits a cold-anomaly maximum in the eastern equatorial region, which tends to be equatorially symmetric and extends westward (Fig. 1b). This reflects the impact of equatorial nonlinear advection[13,20,27].

The two PC time-series display a nonlinear relationship (Fig. 1c). A pIOD event can be constructed by a combination of the two EOFs to realize event diversity. Strong pIOD events appear in the quadrant in which both PCs are positive, that is, with the equatorially symmetric cold anomalies of EOF2 superimposed onto the cold anomalies off Sumatra–Java of EOF1 such that these events are dominated by cold anomalies, as occured in 1997 and 2019 (refs. [20,27]). Strong pIOD events can therefore be described by a strong pIOD index, defined as $S=(\mathrm{PC1}+\mathrm{PC2})/\sqrt{2}$. By contrast, moderate pIOD events appear in the quadrant in which PC1 is positive but PC2 is negative, such that cold anomalies in the east are weak and warm anomalies to the west dominate, as occured in 1982. Moderate pIOD events can thus be described as $M=(\mathrm{PC1}-\mathrm{PC2})/\sqrt{2}$. The two indices separate the observed pIOD events identified by the DMI into two groups: strong events such as 1994, 1997, 2006 and 2019 in the S-index (red curve); and moderate events such as the prominent 1982, 1987 and 2015 events in the M-index (green curve) (Fig. 1c,d). Our approach is similar to a previous analysis on rainfall[20], but we instead use SST, which allows depiction of both strong and moderate pIOD events using SST anomalies. This is the simplest objective description of pIOD diversity compared with other attempts[22,24].

Sea surface temperature anomaly patterns for the two types of pIOD constructed using PC1 and PC2 reproduce the salient features in raw data (Extended Data Figs. 1 and 2), in terms of dominance of the strong cold anomalies in strong pIOD events and by broad-scaled warm anomalies in moderate pIOD events, respectively (Fig. 1e,f). Dynamics that govern the ability to generate the two patterns in part depend on whether the cold anomalies grow northwest towards—and then westward along—the equator, triggering the equatorial nonlinear advective feedbacks that lead to the enhanced cooling represented by PC2, that is, PC2 is a nonlinear function of PC1, which is expressed as $\mathrm{PC2}(t)=\alpha[\mathrm{PC1}(t)]^2+\beta\mathrm{PC1}(t)+\gamma$, similar to that used for describing ENSO diversity[28,29], in which $\alpha$, $\beta$ and $\gamma$ are the nonlinear coefficient, linear coefficient and constant of the quadratic function, respectively. For the observed, obtained from multireanalysis products (see Methods section 'Nonlinear dynamics and model selection'), the mean value of $\alpha$ is 0.42 (Fig. 1c). A large value of $\alpha$ represents strong nonlinearity that distinguishes moderate and strong pIOD events.

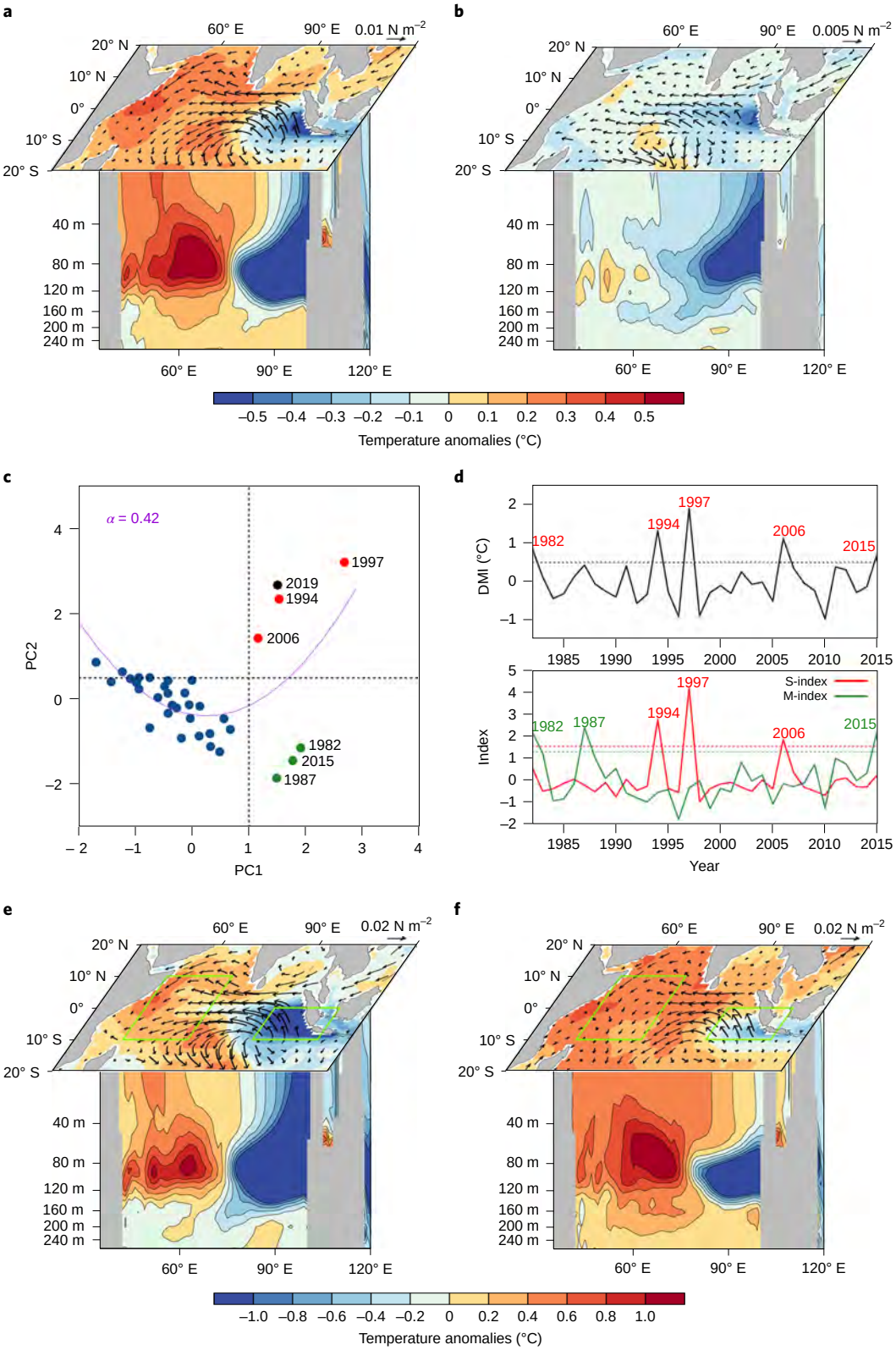
## Simulation of the two pIOD regimes in climate models

We analyse outputs from 35 available CMIP5 (ref. [30]) and 19 CMIP6 (ref. [31]) models by applying the same EOF analysis on SON-averaged SST anomalies from the mean of the first 100 years, quadratically detrended over the full 1900–2099 period (see the 'Data, models and EOF analysis' section in the Methods). These models were forced with historical anthropogenic and natural forcings up to 2005 for CMIP5 or 2014 for CMIP6, and thereafter under a future greenhouse gas emission scenario of representative concentration pathway 8.5 (RCP8.5) for CMIP5 and the equivalent updated

**Fig. 1 | Identification of observed strong and moderate pIOD events. a**, SON SST and tropical vertical potential temperature anomalies (°C, colour) and wind stress anomalies (N m$^{-2}$, vectors) regressed onto the first principal component (PC1) of EOF of SON SST anomalies over the tropical Indian Ocean (5° S to 5° N and 40° E to 100° E) for the 1982–2015 period average across four reanalysis products (see the 'Data, models and EOF analysis' section in the Methods). Subsurface potential temperature and wind stress data are from ORA-S5. The vertical temperature anomalies are calculated as the average over 2.5° S to 2.5° N. **b**, Same as **a**, but for PC2. **c**, The nonlinear relationship between the normalized PC1 and PC2. Red dots indicate strong pIOD years (PC1 > 1 s.d. and PC2 > 0.5 s.d.); green dots denote moderate pIOD years (PC1 > 1 s.d. and PC2 < 0.5 s.d.); the black dot marks the 2019 pIOD, obtained by regressing the SST anomaly pattern over the tropical Indian Ocean in 2019 from the GODAS dataset onto the EOF1 and EOF2 patterns; and the purple curve shows a nonlinear fit $\mathrm{PC2}(t)=\alpha[\mathrm{PC1}(t)]^2+\beta\mathrm{PC1}(t)+\gamma$. **d**, The time-series of a DMI (upper panel) can be split into (red) strong ($S=(\mathrm{PC1}+\mathrm{PC2})/\sqrt{2}$) and (green) moderate pIOD ($M=(\mathrm{PC1}-\mathrm{PC2})/\sqrt{2}$) indices (lower panel) to capture the strong and moderate pIOD events in Fig. 1c. The black dashed line in upper half shows 0.75 s.d. of the DMI, whereas the red and green dashed lines in the lower half show 1.5 s.d. and 1.25 s.d. of the S- and M-index, respectively. **e**, Reconstruction of temperature anomalies (°C, colour shading) and wind stress anomalies (N m$^{-2}$, vectors) of strong pIOD using the two EOFs. The green boxes indicate areas used for calculating the DMI. **f**, Same as **e**, but for moderate pIOD.

scenario for CMIP6, that is, shared socioeconomic pathway 5–8.5 (SSP5–8.5) (refs. [30,31]). The simulated anomaly pattern of either the S- or M-index differs vastly from one model to another (Fig. 2a,b). In contrast to the use of the DMI, our approach accounts for diversity in patterns and thus facilitates assessment of the response of the pIOD SST variability as simulated by each individual model.

Most models examined here simulate a value of $\alpha$ that is lower than the observed and a total of 20 models produce a value of $\alpha$ that is greater than 50% of the observed (Fig. 2c). These 20 models simulate a reasonable nonlinear PC1 and PC2 relationship and the contrast in the cold- and warm-anomaly dominance between the strong and moderate pIOD events (Extended Data Fig. 4 for samples).



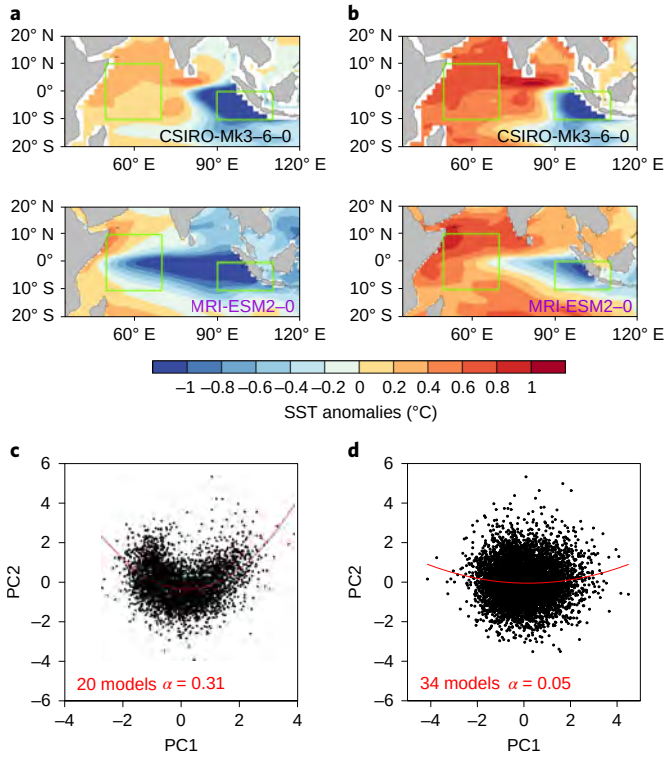



**Fig. 2 | Selection of models based on pIOD nonlinearity. a**, A composite of SST anomalies (in °C) for strong pIOD events (defined as S-index > 1.5 s.d.) in two models, showing vastly different pIOD patterns that cannot be fully represented by the DMI alone. The green boxes indicate areas for calculating the DMI as in ref. [1]. Strong and moderate pIOD indices (represented by the S- and M-index, respectively) are calculated from PCs of the EOF of SON SST anomalies over the tropical Indian Ocean for the period of 1900–2099 for each model. **b**, Same as **a**, but for the composite of moderate pIOD (defined as M-index > 1.25 s.d.) in the two shown models. **c**, A total of 20 models (including the two models shown in Fig. 2a,b) produce a nonlinear relationship between PC1 and PC2 with $\alpha \geqq 0.21$; that is, 50% of the observed value (Fig. 1c), and are thus selected for further analysis. A quadratic fit using PC1 and PC2 time-series from the aggregate of the 20 models produces a value of $\alpha$ of 0.31. **d**, Same as **c**, but for the discarded 34 models with $\alpha < 0.21$.

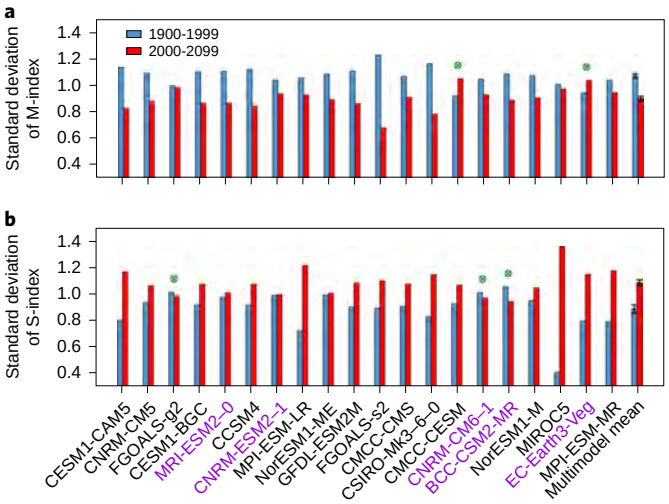



**Fig. 3 | Projected opposite response of moderate and strong pIOD SST.** The result shows a decrease in moderate pIOD variability but an increase in strong pIOD variability. **a**, A total of 18 out of the 20 selected models (90%) generate a reduction in moderate pIOD variability depicted by the M-index from the present-day (1900–1999, blue bars) to future (2000–2099, red bars) climate, with the exception of two models generating an increase (indicated by green circles). The multimodel mean reduction of 16% is statistically significant above the 95% confidence level according to a bootstrap test. The error bars in the multimodel mean represent the 95% confidence level determined by a bootstrap test. **b**, A total of 17 out of the 20 models (85%) simulate an increase in variability of strong pIOD described by the S-index from the present-day to the future climate, with the exception of three models generating a decrease (indicated by green circles). The multimodel mean increase of 22% is statistically significant above the 95% confidence level according to a bootstrap test. Models from CMIP6 are indicated in purple.

We examine this group of models for assessment of possible change in strong and moderate pIOD events under greenhouse warming. The remaining 34 models produce a far weaker nonlinear PC1 – PC2 relationship (Fig. 2d) and are not used.

## Opposite response of the two pIOD regimes to global warming

We compare the standard deviation (s.d.) of the S- and M-index in the present-day (1900–1999) and future (2000–2099) periods. There is an opposite response: a total of 18 out of 20 (that is, 90%) models simulate decreased variability of the M-index (Fig. 3a), whereas a total 17 out of 20 models generate increased variability of the S-index in the future period (red bars, Fig. 3a,b). The ensemble-mean is a decrease of 16% for moderate, but an increase of 22% for strong pIOD variability, each statistically significant above the 95% confidence according to a bootstrap test (see the 'Statistical test' section in the Methods). This is in sharp contrast to the result using the DMI, which shows no statistically significant change in the multimodel mean change (Extended Data Fig. 5).

The decrease in M-index variability translates to a 52% reduction in occurrences of moderate pIOD events with an M-index > 1.25 s.d., as supported by 85% of the selected models (Extended Data Fig. 6a); however, the increase in S-index variability translates to a 66% increase in occurrences of strong pIOD events with an S-index > 1.5 s.d. (Extended Data Fig. 6b)—from one event every 13.3 years in the present-day climate to one event every eight years in the future climate, as supported by 80% of the selected models. The opposite response is despite a lack of frequency change in strong and moderate pIOD events combined (Extended Data Fig. 6a,b), in contrast to a tendency of increased pIOD frequency using DMI (ref. [32]). These results hold using other threshold values (Extended Data Fig. 6c,d). Climate extremes associated with strong pIOD events are thus likely to occur more frequently under greenhouse warming. The 2019 strong pIOD happened to occur in conjunction with a moderate Pacific warm condition[33], leading to more extreme impacts.

Using 30 models simulating a value of $\alpha$ that is at least 33% of the observed value, the intermodel consensus on the change of the two indices is still strong. A total of 27 out of 30 models (90%) simulate a decrease in moderate pIOD variability and a total of 23 out of 30 models (77%) produce an increase in strong pIOD variability, and their ensemble-mean changes are statistically significant (Extended Data Fig. 7). Sensitivity to emission scenarios shows that under RCP4.5, 13 out of 15 (87%) CMIP5 models that simulate a value of $\alpha$ greater than 50% of the observed value generate an increase in S-index variability, and 93% of the models produce a reduction in M-index variability (Extended Data Fig. 8).

## Associated mechanisms arising from mean state changes

There is no intermodel relationship between changes in variability or frequency of a strong pIOD event and changes in El Niño

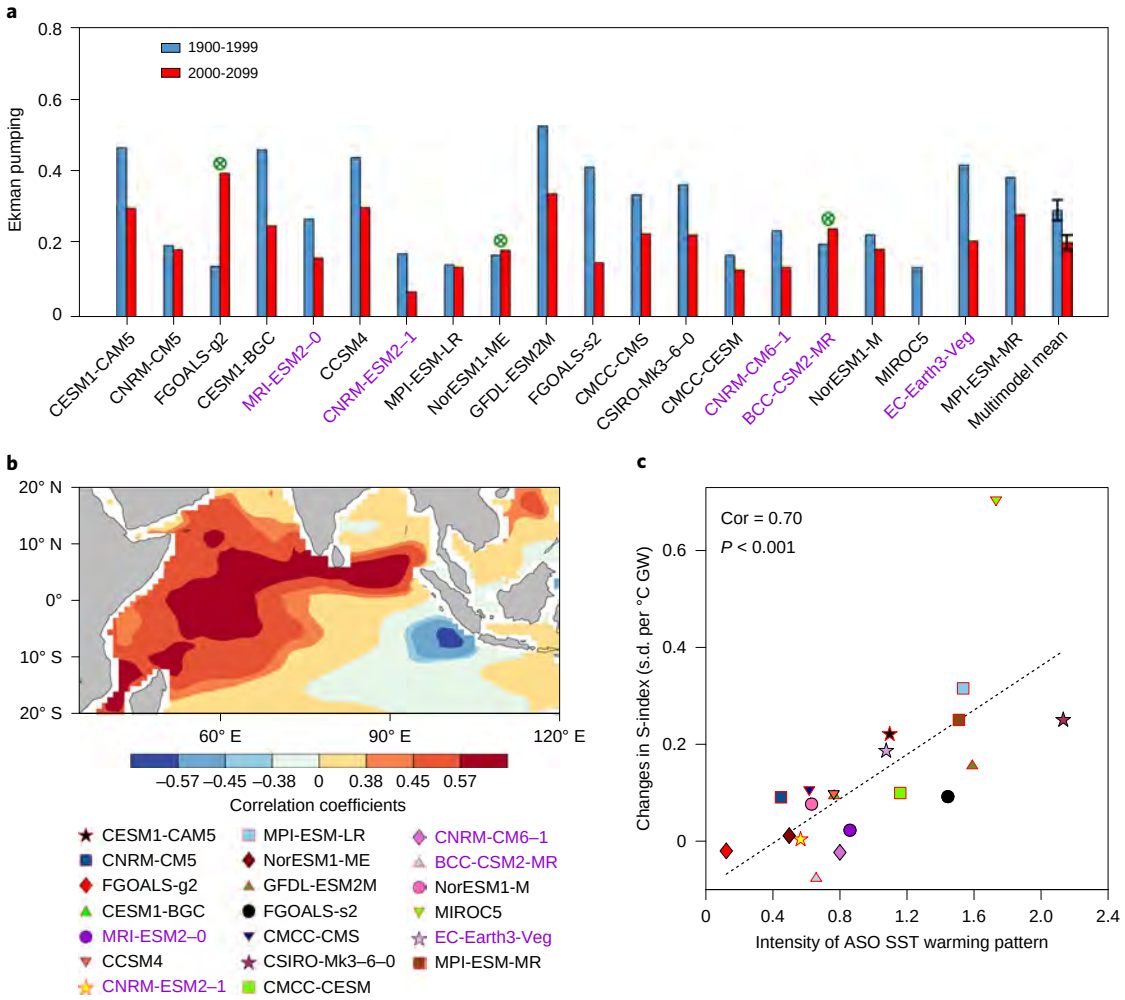



**Fig. 4 | Mechanisms for the projected opposite change in moderate and strong pIOD. a**, Ekman pumping term during the development phase (August, September, October (ASO)) averaged over the southwestern tropical Indian Ocean (10° S to 0° and 50° E to 80° E) regressed onto the M-index over the present-day (1900–1999, blue bars) and future (2000–2099, red bars) climate in the 20 selected models. A total of 17 out of the 20 selected models (85%) simulate decreased Ekman pumping during future moderate pIOD events. Models that simulate an increase are indicated by green circles. The associated multimodel mean decrease of 30% is statistically significant above the 95% confidence level according to a bootstrap test. Models from CMIP6 are indicated in purple. A time-series of the Ekman pumping term is normalized before regression. The error bar in the multimodel mean represents the 95% confidence level determined by a bootstrap test. **b**, A map of the intermodel correlation coefficients between the ASO warming pattern and the change in the S-index variability. The warming pattern is defined as the change in SST between the future and present-day. The S-index variability change and warming pattern have been scaled by the corresponding increase in global mean SST in each model. Correlation coefficients of ±0.38, ±0.45 and ±0.57 are significant at the 90%, 95% and 99% confidence levels, respectively. **c**, The statistically significant relationship between the intensity of ASO SST warming pattern and the change in s.d. of the S-index is shown, with the correlation (Cor) and its *P*-value indicated. The intensity of the SST warming pattern over 20° S to 20° N and 40° E to 100° E in each model is obtained by regressing onto the multimodel mean SST warming pattern in the same domain; the S-index variability change and warming pattern have been scaled by the corresponding increase in global mean SST in each model, taken as the intensity of global warming (GW).

variability or frequency (Extended Data Fig. 9a,c), which suggests that the changes of strong pIOD are not due to changes in the El Niño. There is a relationship between changes in variability or frequency of a moderate pIOD with changes in El Niño, but the fact that in the majority of models there is an increase in El Niño variability and El Niño frequency in contrast to a decrease in moderate pIOD (Extended Data Fig. 9b,d) rules out changes in the El Niño as the cause.

Instead, the opposite response arises from the mean state changes, featuring faster warming in the western and northern than the eastern and southern tropical Indian Ocean, supported by a trend of equatorial southeasterlies[15–19] and enhanced evaporation in the west (Extended Data Fig. 10). Under greenhouse warming, the lower troposphere warms faster than the surface, such that the response of equatorial easterlies to SST anomalies weakens[16]. Ekman pumping, which is the main forcing for warm anomalies in the tropical western Indian Ocean for moderate pIOD events (Extended Data Fig. 3), weakens in the future climate (Fig. 4a), causing decreased M-index variability. For cold-anomaly-dominated pIOD events (that is, strong pIOD events), the faster mean warming west and northwest of the tropical Indian Ocean[18,19] is conducive to anomalous atmosphere convection there, facilitating anomalous easterlies that extend to the equatorial central and western Indian Ocean, leading to a shallowing thermocline, enhanced equatorial nonlinear zonal and vertical advection, which is in turn conducive to cold anomalies[13,20,27]. Intermodel relationships between warming

trends and changes in S-index variability show that a greater intensity of the warming pattern leads to a great increase in S-index variability (Fig. 4b,c).

## Conclusions

Our finding of an opposite response to greenhouse warming—with increased SST variability of strong pIOD events but decreased SST variability of moderate pIOD events—is in stark contrast with previous findings of no consensus using the DMI. Our robust result arises from a separation of the two types of pIOD events that are governed by different dynamics, which provides a physical basis for selecting models used in our pIOD future projections. This projected change of a reduced frequency of moderate pIOD but with more occurrences of stronger pIOD events has far-reaching implications for climate and weather extremes over regions influenced by the pIOD. For example, severe floods in eastern African regions and devastating bushfires in Australia, as seen in 2019, are likely to occur more frequently.

## Online content

Any methods, additional references, Nature Research reporting summaries, source data, extended data, supplementary information, acknowledgements, peer review information; details of author contributions and competing interests; and statements of data and code availability are available at https://doi.org/10.1038/s41558-020-00943-1.

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## Methods

**Data, models and EOF analysis.** We use four SST reanalysis products to construct pIOD patterns, covering a common period from 1982 to 2015. Most of these products are produced in a system that specifically considers the impact of subsurface ocean temperature on SST, which is essential for the nonlinear oceanic feedback responsible for the pIOD dominated by cold SST anomalies, that is, strong pIOD episodes. The only exception is OISST, which although not derived from a system explicitly with a subsurface ocean, assimilates weekly observations from different platforms including satellites, ships and buoys. For each of these four data, anomalies are constructed with reference to the mean of the full period, and then detrended. A multiproduct average of the anomalies is then calculated. These products are:

- GODAS (NCEP Global Ocean Data Assimilation System from 1980 to 2019) (ref. [34]);
- ORA-S5 (ECMWF Ocean Reanalysis System 5 from 1979 to 2018) (ref. [35]);
- OISST v2 (NOAA Optimum Interpolation SST version 2 from 1982 to 2019) (ref. [36]);
- SODA3.3.1 (Simple Ocean Data Assimilation version 3.3.1 from 1980 to 2015) (ref. [37]).

To assess the impact of greenhouse warming on pIOD changes, we take outputs from 35 CMIP5 (ref. [30]) and 19 CMIP6 (ref. [31]) coupled global climate models forced with historical anthropogenic and natural forcings, and future greenhouse gases under RCP8.5 for CMIP5 and SSP5–8.5 for CMIP6, covering a 200-year period of 1900–2099 (refs. [30,31]). Monthly anomalies referenced to the climatology of 1900–1999 were constructed and quadratically detrended.

We apply EOF analysis[38] to the SON average of SST anomalies in an equatorial Indian Ocean domain (5° S to 5° N, 40° E to 100° E), which yield the two dominant modes, each with a principal pattern and a PC scaled to unity. The two PC time-series display a nonlinear relationship. A combination of the two EOFs is used to describe the two types of pIOD. Indices of moderate and strong pIOD are described by the M-index $((\mathrm{PC1}-\mathrm{PC2})/\sqrt{2})$ and the S-index $((\mathrm{PC1}+\mathrm{PC2})/\sqrt{2})$, respectively. The two indices separate the pIOD events identified by the DMI into two groups: strong events such as 1994, 1997 and 2006 in the S-index (red curve), and moderate events such as 1982, 1987 and 2015 events (green curve) in the M-index (Fig. 1d). Using these ocean-model-based products identifies the 2006 event as a strong pIOD, as opposed to a moderate one using non-oceanic-based products[20]. Data for the 2019 pIOD event are available in GODAS (ref. [34]), and a pattern regression onto EOF1 and EOF2 shows it to be a strong pIOD, second only to the 1997 event (Fig. 1c).

**Heat budget analysis.** We examine the surface heat balance of the tropical Indian Ocean, which is expressed as:

$$\partial T^a/\partial t = -\begin{bmatrix} \left(u^a\partial T^a/\partial x + \overline{u}\partial T^a/\partial x + u^a\partial \overline{T}/\partial x\right) \\ +\left(v^a\partial T^a/\partial y + \overline{v}\partial T^a/\partial y + v^a\partial \overline{T}/\partial y\right) \\ +\left(w^a\partial T^a/\partial z + \overline{w}\partial T^a/\partial z + w^a\partial \overline{T}/\partial z\right) \end{bmatrix} + Q + \text{residual} \qquad (1)$$

The variables $T$, $u$, $v$ and $w$ are the potential temperature and the zonal, meridional and vertical ocean current velocities, respectively, averaged over the top 50 m (vertical motion over the top 50 m is represented by $w$ at 50 m). The vertical gradient of climatological temperature $(d\overline{T}/dz)$ is taken as the difference between the average over the top 50 m and temperature at 60 m depth. Differential operators, $x$, $y$, $z$, and $t$ represent zonal, meridional and vertical directions, and time, respectively. All variables are from the ECMWF ORA-S5 dataset[35] at a horizontal resolution of 1° latitude by 1° longitude for the period of 1979 to 2018. The rate of change of the mixed-layer temperature $(d\overline{T}/dt)$ is calculated using a centred-difference approximation. The superscript $a$ and overbar denote anomalous and long-term averaged quantities, respectively. The tendency of the surface temperature is balanced by zonal advection of heat by the zonal currents (first bracketed terms on the right-hand side), meridional advection (second bracketed terms), vertical advection (third bracketed terms), the net surface air–sea heat flux ($Q$) and all other factors not explicitly expressed (residual), such as mixing and diffusion. We examine the heat budget terms averaged over the developing period of IOD events (August to October), which contributes most to the mature SST anomalies. All variables in (1) are linearly detrended.

We first examine the heat budget for the observations in an area average over the equatorial region (10° S to 10° N and 40° E to 100 °E). For strong pIOD events, nonlinear zonal and vertical advection terms are particularly important for generating cold anomalies in the equatorial eastern Indian Ocean[20], and Ekman pumping plays a key role in the generation of warm anomalies (Extended Data Fig. 3a). For moderate pIOD events, Ekman pumping dominates warming (Extended Data Fig. 3a). Inspection of $d\overline{T}/dt$ finds that the region where maximum warming tendency occurs is in the southwestern tropical Indian Ocean (10° S to 0°, 50° E to 80° E, green box of Extended Data Fig. 3b,c). Propagation into the region of downwelling Rossby waves driven by equatorial easterly anomalies associated with a developing pIOD, and their subsequent reflection at the western boundary as downwelling equatorial Kelvin waves induce the broad surface warming[26]. That Ekman pumping in the southwestern tropical Indian Ocean is an important term in the total tendency, and is driven by equatorial easterly anomalies, is reaffirmed by a simple analysis (Extended Data Fig. 3d–f).

We then examine the heat budget for climate models. Data from all models have been interpolated to a horizontal grid of 1°×1° (bilinear interpolation) and a vertical grid with 5 m intervals (linear interpolation). The vertical velocity $w$ at 50 m is used to represent the vertical motion over the upper 50 m layer. Monthly anomalies of $w$ and zonal wind stress are calculated with reference to the 1900–1999 climatology and quadratically detrended. As in observations, nonlinear equatorial zonal and vertical advection are dominant in forcing strong pIOD events, and these terms increase under greenhouse warming[20].

For the present-day and future model climate, we take the climatological mean temperature $(\overline{T})$ as the average over the 1900–1999 and 2000–2099 periods, respectively. There is a general decrease in the response of zonal wind stress to SST in the equatorial eastern Indian Ocean, as a consequence of a more stable atmosphere that results from a faster warming in the lower troposphere than at the surface under greenhouse warming[16]. The associated decrease in variability of zonal wind stress anomalies, which drive Ekman pumping that dominates moderate pIOD events, leads to decreased variability of moderate pIOD in the future climate. In the 20 selected models, 19 models generate a future decrease of $w$ variability in the development phase over the southwestern tropical Indian Ocean (green box, Extended Data Fig. 3b). A bootstrap test finds that the multimodel mean decrease of 21% is statistically significant. By contrast, for strong pIOD events, equatorial nonlinear zonal and vertical advection occurs more frequently as a consequence of faster warming in the northwestern equatorial Indian Ocean under greenhouse warming[20].

**Nonlinear dynamics and model selection.** During strong pIOD events, the initial cooling grows such that it extends northward first and then westward along the equator, triggering the equatorial nonlinear ocean–atmosphere feedback: an equatorial west-minus-east zonal SST gradient drives easterly anomalies, leading to nonlinear vertical and nonlinear zonal advection extending the cold anomalies along the equatorial eastern Indian Ocean[13,20,27], as represented by PC2. The dynamics that govern PC2 and the associated pattern depends nonlinearly on cold anomalies presented in PC1 that first grow northwest towards and then westward along the equator, such that $\mathrm{PC2}(t)=\alpha[\mathrm{PC1}(t)]^2+\beta\mathrm{PC1}(t)+\gamma$. For the observed, the value of $\alpha$ is 0.42 (Fig. 1c); a greater value of $\alpha$ means a clearer separation of the two types of the pIOD events.

Not all models produce the nonlinear dynamics that separate the strong and moderate pIOD events, and most models produce a weaker nonlinearity than seen in the observed. We select 20 CMIP5 and CMIP6 models that simulate an $\alpha$, calculated over the 200 years, greater than half of the observed value. The 20 selected models together yield a mean $\alpha$ amplitude of 0.31 (Fig. 2c). A total of 30 models simulate an $\alpha$ greater than 33% of the observed value and we use this group to test the sensitivity of our results.

**Statistical significance test.** A bootstrap method[39] is used to examine whether the increased S-index variance and decreased M-index variance are statistically significant. The 20 s.d. values of the S- and M-index in the present-day climate (1900–1999) from the 20 selected models are resampled randomly to construct 10,000 realizations of the mean standard deviation. In this random resampling process, a model is allowed to be selected again. The s.d. of the 10,000 interrealizations of mean s.d. in the present-day climate is 0.0312 for the S-index (0.0157 for the M-index). The same test is carried out for the future climate (2000–2099) and the s.d. of the interrealization is 0.0212 for the S-index (0.0185 for the M-index). The increased s.d. for the S-index (or decreased s.d. for the M-index) in the future period is greater than the sum of the two s.d. values, indicating statistical significance above the 95% confidence level.

## Data availability

Data relevant to the paper can be downloaded from websites listed below:
OISST v2 at https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.html;
ORA-S5 at https://icdc.cen.uni-hamburg.de/daten/reanalysis-ocean/easy-init-ocean/ecmwf-oras5.html;
SODA3.3.1 at https://www2.atmos.umd.edu/~ocean/index_files/soda3.3.1_mn_download.htm;
GODAS at https://www.esrl.noaa.gov/psd/data/gridded/data.godas.html;
CMIP5 database at http://www.ipcc-data.org/sim/gcm_monthly/AR5/;
CMIP6 database at https://esgf-node.llnl.gov/projects/cmip6/;
19 CMIP6 models are used in this study, including: MRI-ESM2–0 (refs. [40,41]); CNRM-ESM2–1 (refs. [42,43]); CNRM-CM6–1 (refs. [44,45]); BCC-CSM2-MR (refs. [46,47]); EC-Earth3-Veg (refs. [48,49]); NorESM2-LM (refs. [50,51]); MIROC6 (refs. [52,53]); CESM2 (refs. [54,55]); CAMS-CSM1-0 (refs. [56,57]); CESM2-WACCM (refs. [58,59]); MIROC-ES2L (refs. [60,61]); IPSL-CM6A-LR (refs. [62,63]); INM-CM4-8 (refs. [64,65]); INM-CM5-0 (refs. [66,67]); CanESM5 (refs. [68,69]); MPI-ESM1-2-LR (refs. [70,71]); UKESM1-0-LL (refs. [72,73]); MCM-UA-1-0 (refs. [74,75]); GFDL-CM4 (refs. [76,77]).

## Code availability

Codes for calculating EOF, the parameter $|\alpha|$ can be downloaded from https://drive.google.com/open?id=1d2R8wKpFNW-vMIfoJsbqIGPIBd9Z_8rj.

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## Author contributions

W.C. conceived the study and wrote the initial manuscript in discussion with K.Y. K.Y. performed all analysis and generated final figures. All authors contributed to interpreting results, discussion of the associated dynamics and improvement of this paper.

## Competing interests

The authors declare no competing interests.

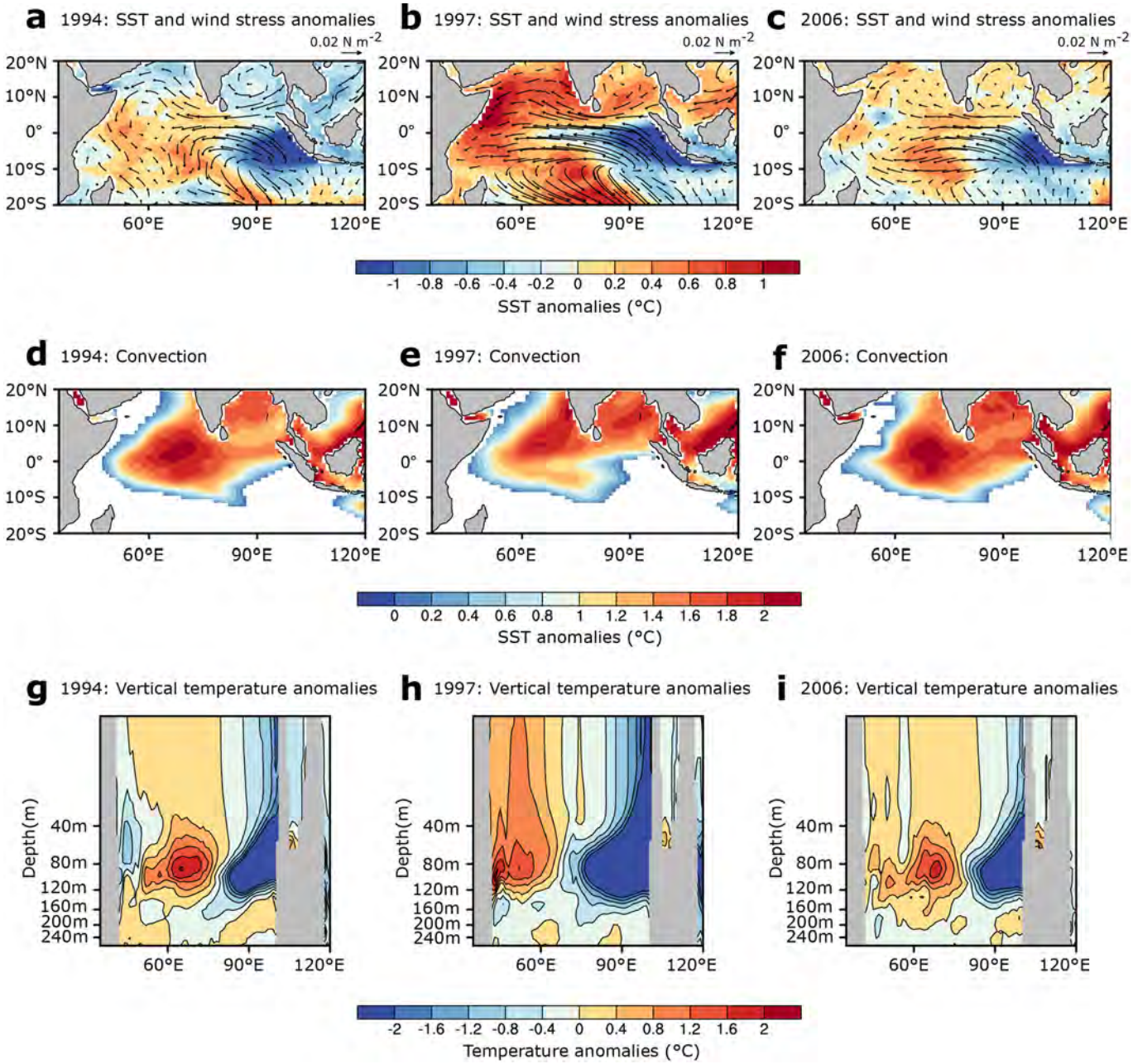
## Additional information

**Extended data** is available for this paper at https://doi.org/10.1038/s41558-020-00943-1.

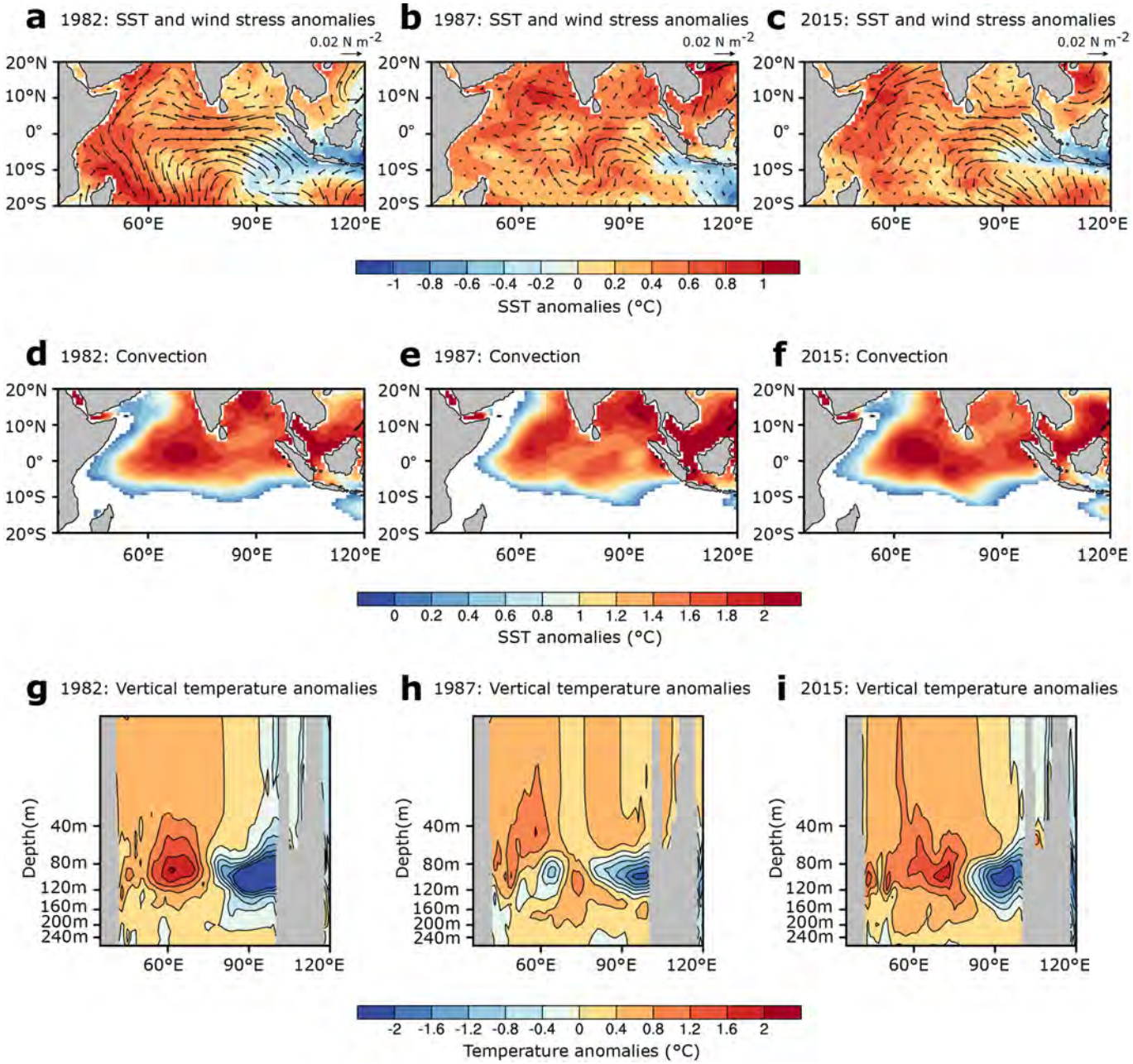
**Correspondence and requests for materials** should be addressed to W.C. or L.W.

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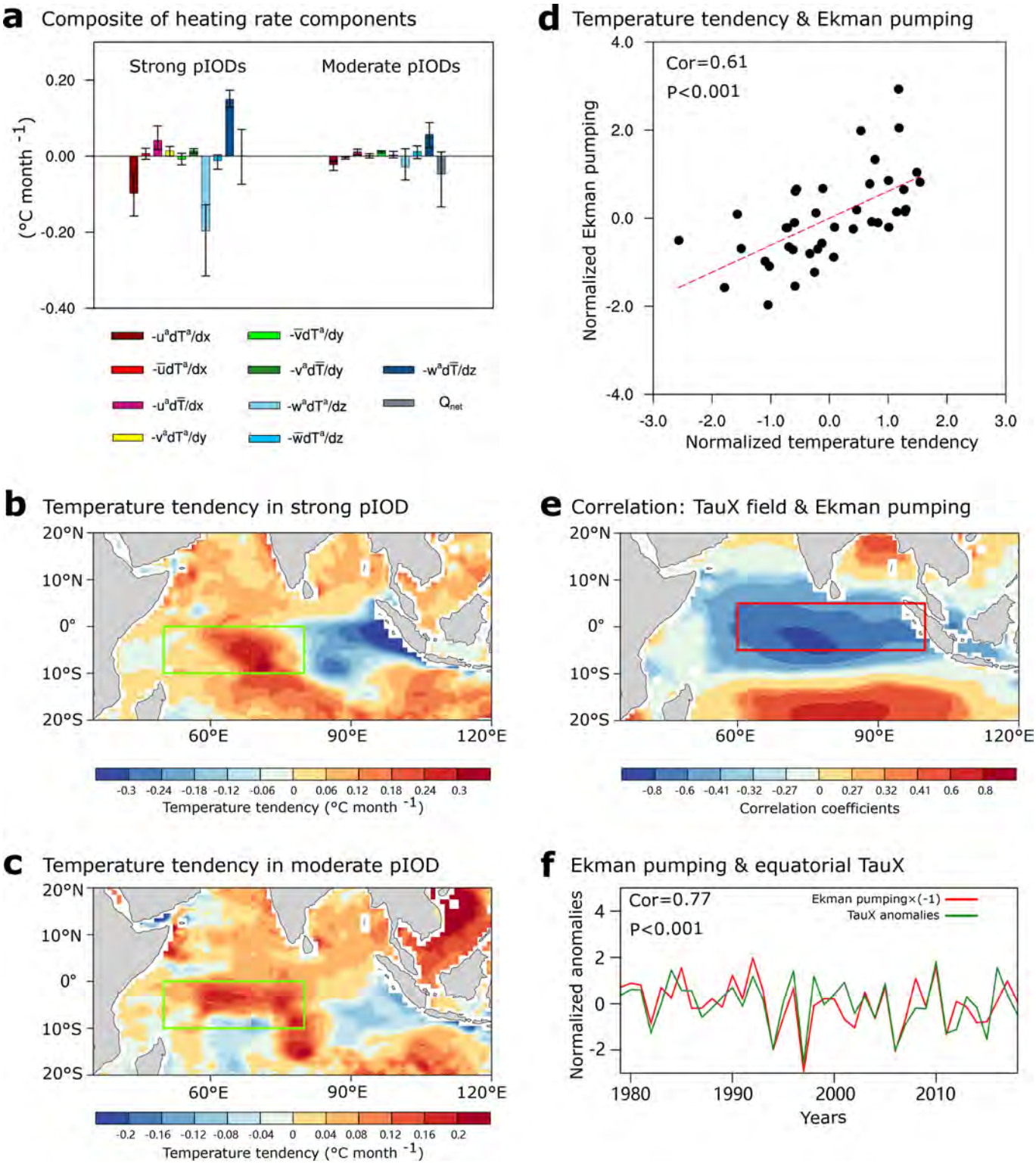
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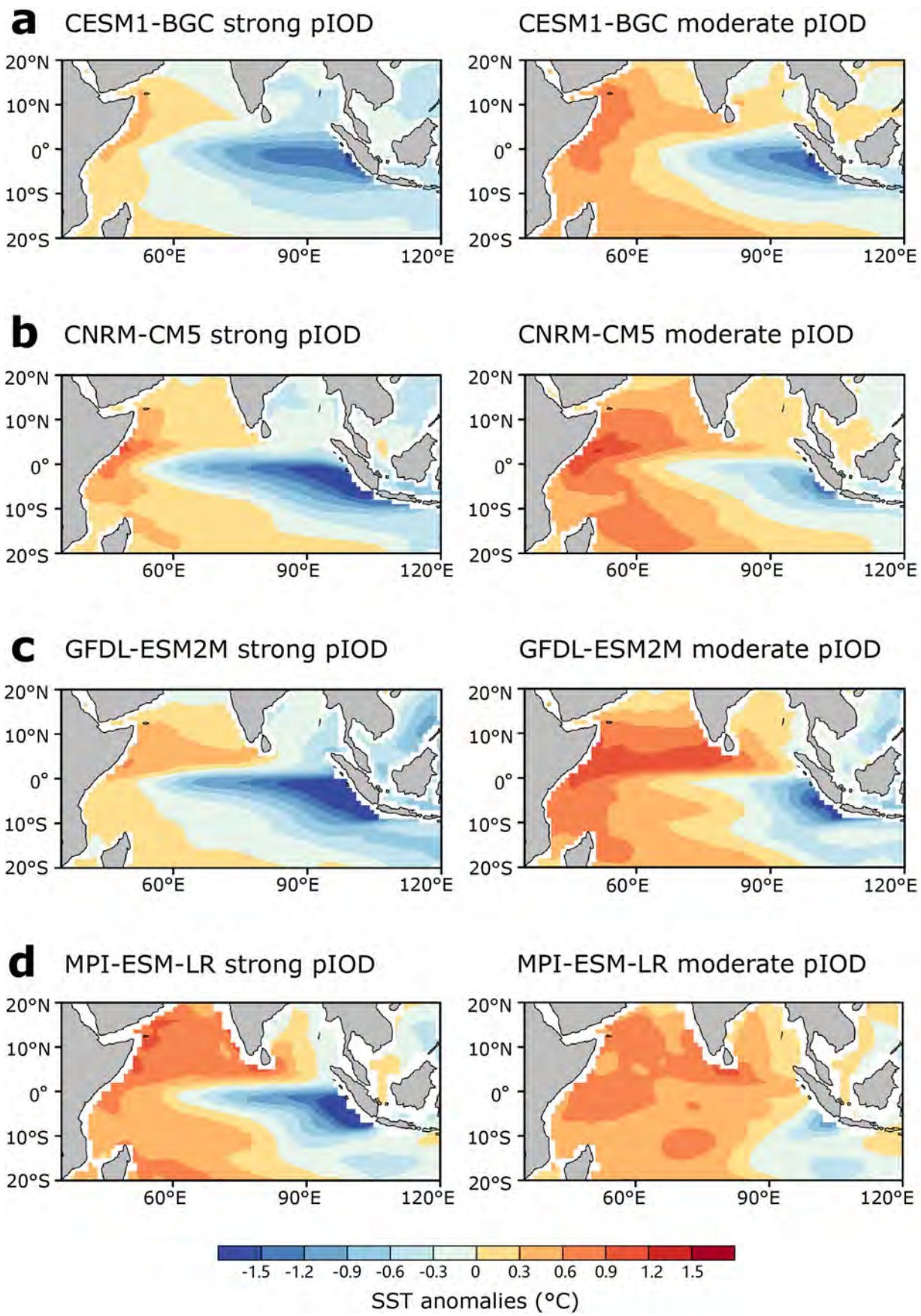

**Extended Data Fig. 1 | Characteristics of strong pIOD events. a–c**, SON SST anomalies (°C, color shading) and wind stress anomalies (N $m^{-2}$, vectors) for the 1994, 1997, and 2006 strong pIOD event, respectively. The pattern is dominated by strong cooling in the equatorial eastern Indian Ocean. Strong equatorial easterly anomalies extend to the central-western Indian Ocean. The anomalies are referenced to the average over the 1982–2015 period. **d–f**, SON SST anomalies (in °C) relative to the SST threshold for tropical convection, defined as the SON SST average over 20°S to 20°N, for the 1994, 1997, and 2006 strong pIOD event, respectively. Areas with SST lower than the convection threshold have been masked; there is a large area of suppressed convection in the equatorial eastern Indian Ocean. **g–i**, Equatorial vertical potential temperature anomalies (in °C) over SON of 1994, 1997, and 2006, respectively. The equatorial vertical potential temperature is calculated as the average between 2.5°S to 2.5°N. Strong equatorial cooling is seen in the eastern Indian Ocean. Vertical potential temperature and wind stress data are from ORA-S5.



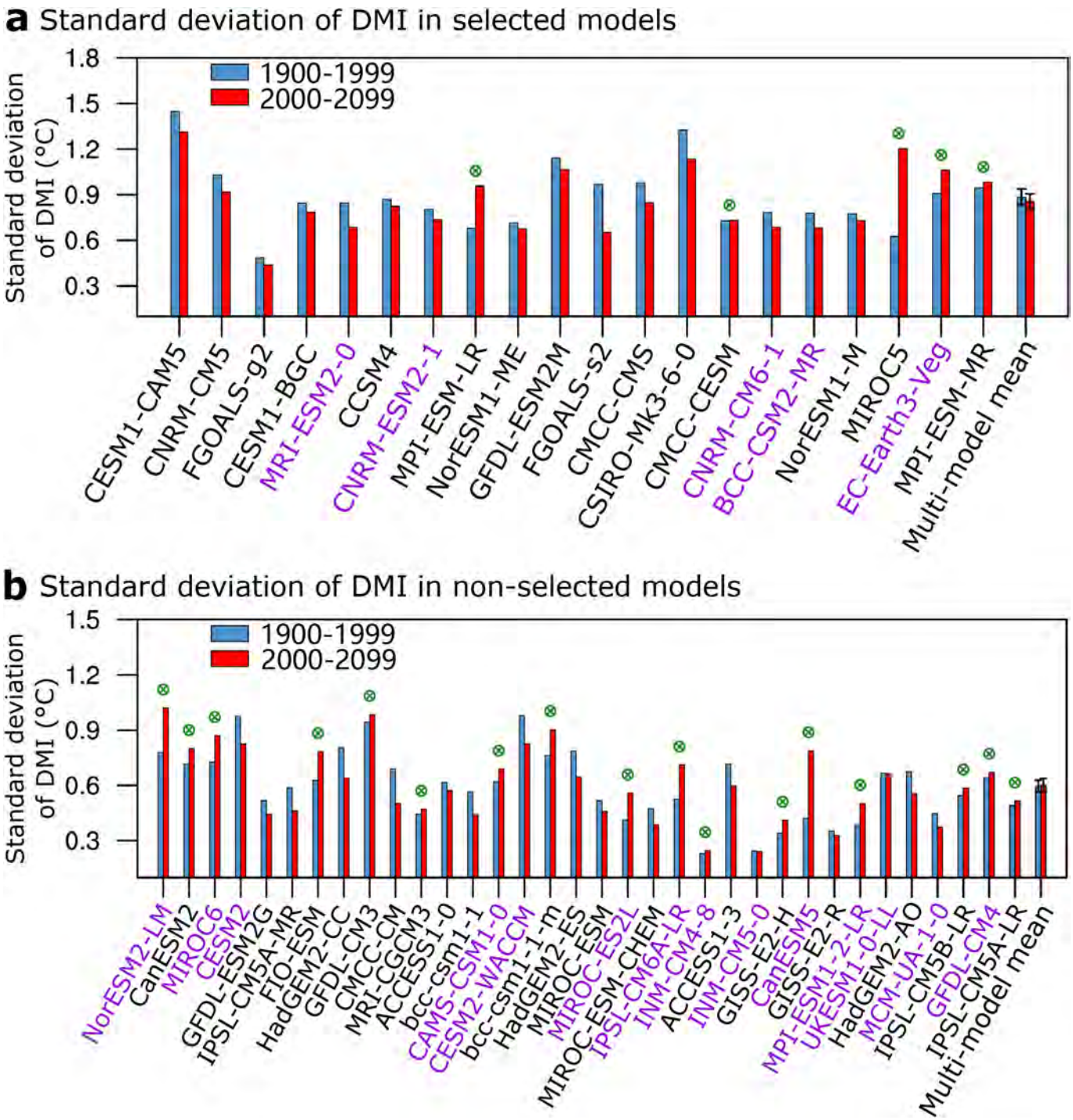

**Extended Data Fig. 2 | Characteristics of moderate pIOD events. a–c**, SON SST anomalies (°C, color shading) and wind stress anomalies (N $m^{-2}$, vectors) for the 1982, 1987, and 2015 moderate pIOD event, respectively. The pattern is dominated by a broad-scale warming over the western Indian Ocean. Easterly anomalies are confined to the equatorial central-eastern Indian Ocean. The anomalies are referenced to the average over the 1982–2015 period. **d–f**, SON SST anomalies (in °C) relative to SST threshold for tropical convection, defined as SON SST average over 20°S to 20°N for 1982, 1987, and 2015 moderate pIOD event, respectively. Areas with SST lower than the convection threshold have been masked; there is a small area of suppressed convection in the equatorial southeastern Indian Ocean. **g–i**, Equatorial vertical potential temperature anomalies (in °C) over SON of 1982, 1987, and 2015, respectively. The equatorial vertical potential temperature is calculated as the average between 2.5°S to 2.5°N. There is a general lack of cooling in the equatorial eastern Indian Ocean as cooling occurs only off Sumatra-Java. Vertical potential temperature and wind stress data are from ORA-S5.



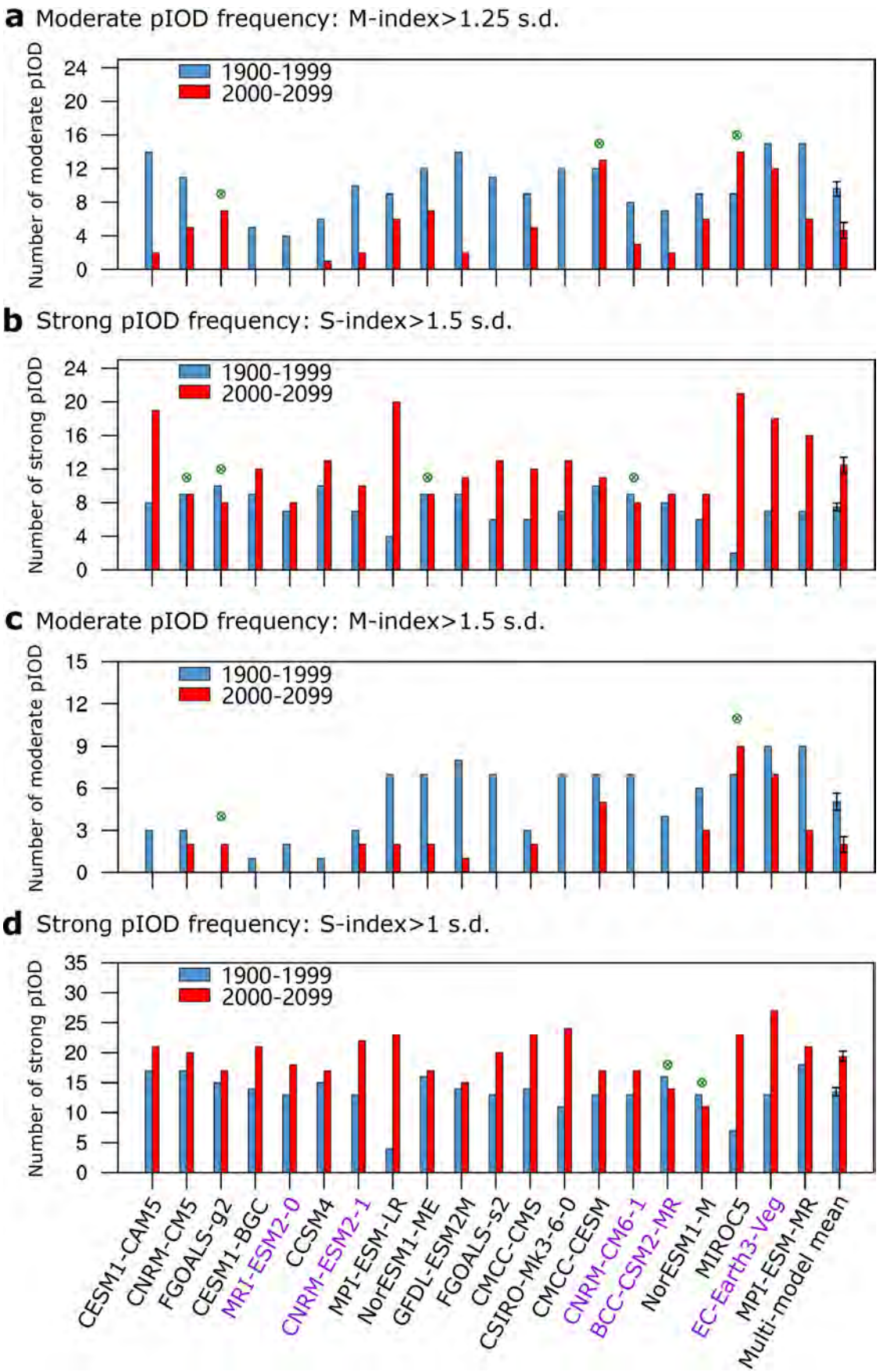

**Extended Data Fig. 3 | Heat budget analysis based on observations. a**, Composite of heat budget terms over the equatorial Indian Ocean (10°S–10°N, 40°E–100°E) in August-September-October (ASO). The uncertainty bar on each composite represents the range over the three strong or three moderate pIOD events. **b**, **c**, Composite of ASO temperature tendency during strong and moderate pIOD events. The green box marks the area where the main warm temperature tendency is located over the southwestern Indian Ocean (10°S–0°, 50°E–80°E). **d**, Relationship between normalized temperature tendency and the Ekman pumping term in ASO averaged over the main warming area in the southwestern Indian ocean (10°S–0°, 50°E–80°E, green box in c), with statistical properties shown. **e**, Map of ASO correlation coefficients between the Ekman pumping term averaged over 10°S–0°, 50°E–80°E and zonal wind stress (TauX) anomalies. Correlation coefficients of ±0.27, ±0.32, ±0.41 are statistically significant at the 90%, 95%, and 99% confidence levels, respectively. The area indicated by the red box is used to calculate the average zonal wind stress anomalies. **f**, Time series of normalized ASO Ekman pumping term averaged over the western Indian Ocean (10°S–0°, 50°E–80°E) (red curve) and zonal wind stress (TauX) anomalies averaged over the equatorial Indian Ocean (5°S–5°N, 60°E–100°E, red box in e) (green curve). The Ekman pumping term has been multiplied by −1 for plotting. The data used in this figure are from ORA-S5 for the period of 1979–2018.




**Extended Data Fig. 4 | Anomaly patterns of strong and moderate pIOD in some selected models.** A composite of SON SST anomalies (°C) for strong pIOD with an S-index>1.5 s.d. and moderate pIOD with an M-index>1.25 s.d. in **a** CESM1-BGC, **b** CNRM-CM5, **c** GFDL-ESM2M, **d** MPI-ESM-LR. Left panels show strong pIOD and right panels show moderate pIOD. These show that anomaly patterns can be rather different from one model to another.

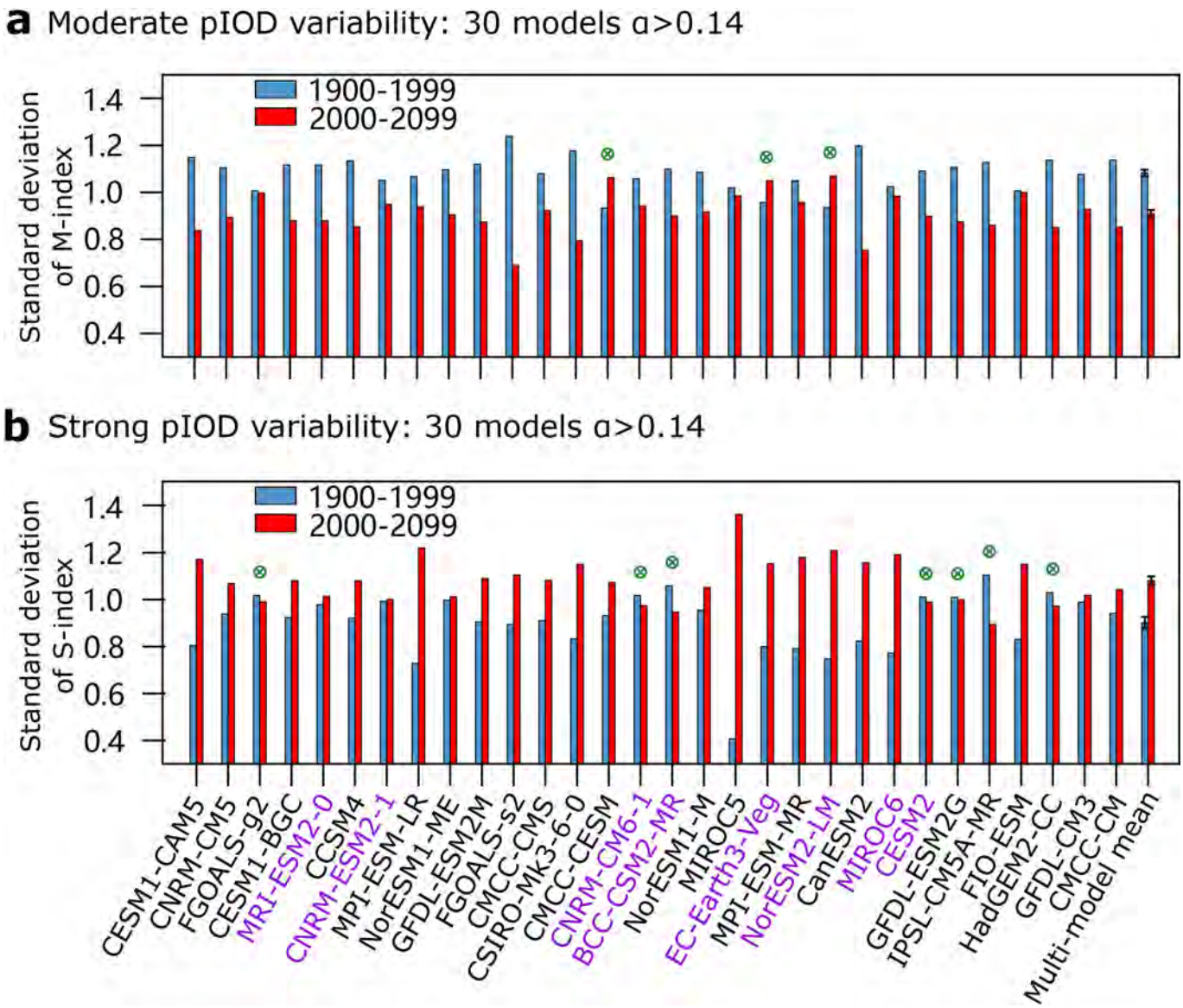



**Extended Data Fig. 5 | Uncertainty in projected pIOD SST using the DMI. a**, Comparison of the standard deviation of the DMI in the present-day (1900–1999) and future (2000–2099) climate in 20 models that simulate a nonlinear coefficient α greater than 50% of the observed in the quadratic fit between the first two principal components (PC1 and PC2) from EOF analysis on SON SST anomalies, $PC2(t) = \alpha\,[PC1(t)]^2 + \beta\,PC1(t) + \gamma$. Although a total of 15 out of the 20 selected models simulate lower DMI variability in the future period (red bars) than in the present-day climate (blue bars), the multi-model mean decrease is not statistically significant. The error bar represents the 95% confidence level. Models that simulate an increase in variability are indicated by green circles. Models from CMIP6 are shown in purple. **b**, Same as **a**, but for 34 models that simulate an α smaller than 50% of the observed value. A total of 17 out of the 34 models (50%) simulate smaller variability in the DMI in the future period than in the present-day period. There is no inter-model consensus, and the multi-model mean change is not statistically significant.

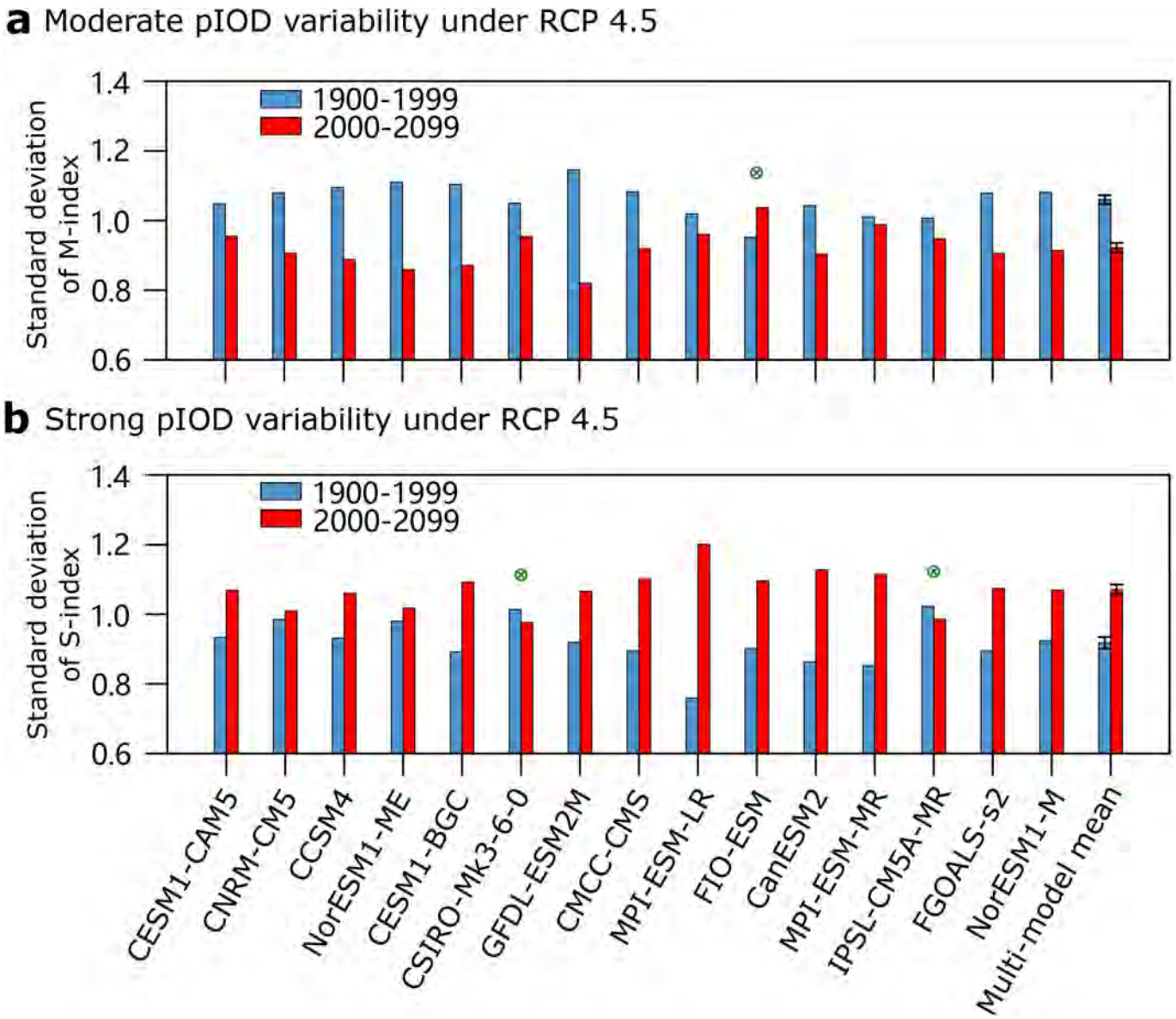



**Extended Data Fig. 6 |** See next page for caption.

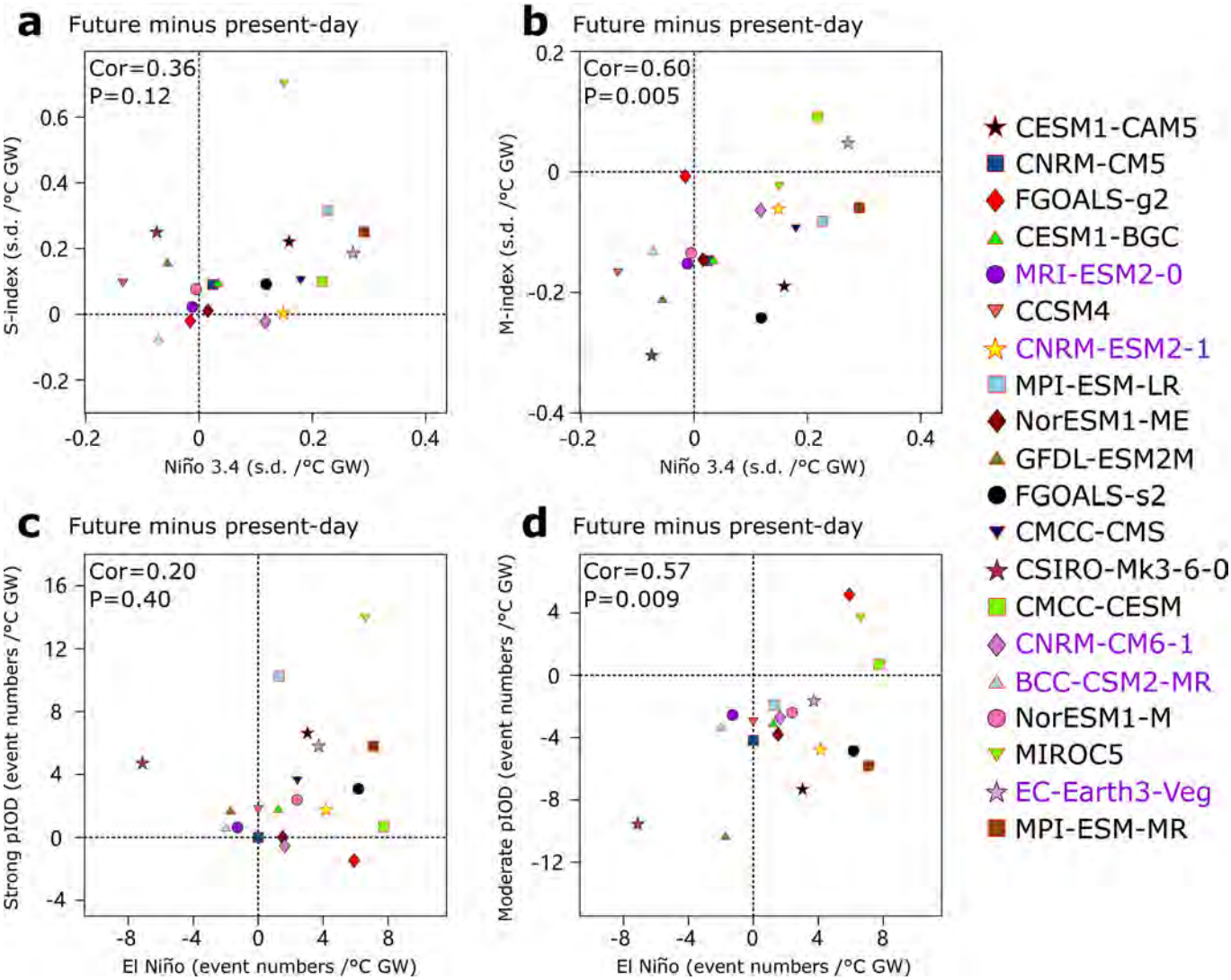
**Extended Data Fig. 6 | Projected change in pIOD frequency. a**, Moderate and **b**, strong pIOD using a threshold of M-index>1.25 s.d. and S-index>1.5 s.d., respectively. Shown is the comparison of frequency (events per 100 years) in the present-day (1900–1999) and future (2000–2099) climate in the 20 selected models, that is, with an α greater than 50% of the observed value. A total of 17 out of the 20 models (85%) simulate a decreased frequency in moderate pIOD events, with a statistically significant decrease of 52% in the multi-model mean, from 9.60 events per 100 years in the present-day (blue bars) to 4.65 events per 100 years (red bars) in the future climate. By contrast, a total of 16 out of the 20 models (80%) simulate an increased frequency in strong pIOD events, with a statistically significant increase of 66% in the multi-model average, from 7.5 events per 100 years in the present-day (blue bars) to 12.45 events per 100 years (red bars) in the future climate. Models that simulate an opposite change to the multi-model mean are indicated by green circles. **c**, Same as **a**, but using a threshold of M-index>1.5 s.d. A total of 18 out of the 20 selected models (90%) simulate a decreased frequency in moderate pIOD events, with a statistically significant decrease of 60% in the multi-model ensemble mean. **d**, Same as **b**, but using a threshold of S-index>1 s.d. A total of 18 out of the 20 models (90%) simulate an increased frequency in strong pIOD, with a statistically significant increase of 44% in the multi-model ensemble mean. Models from CMIP6 are indicated in purple.



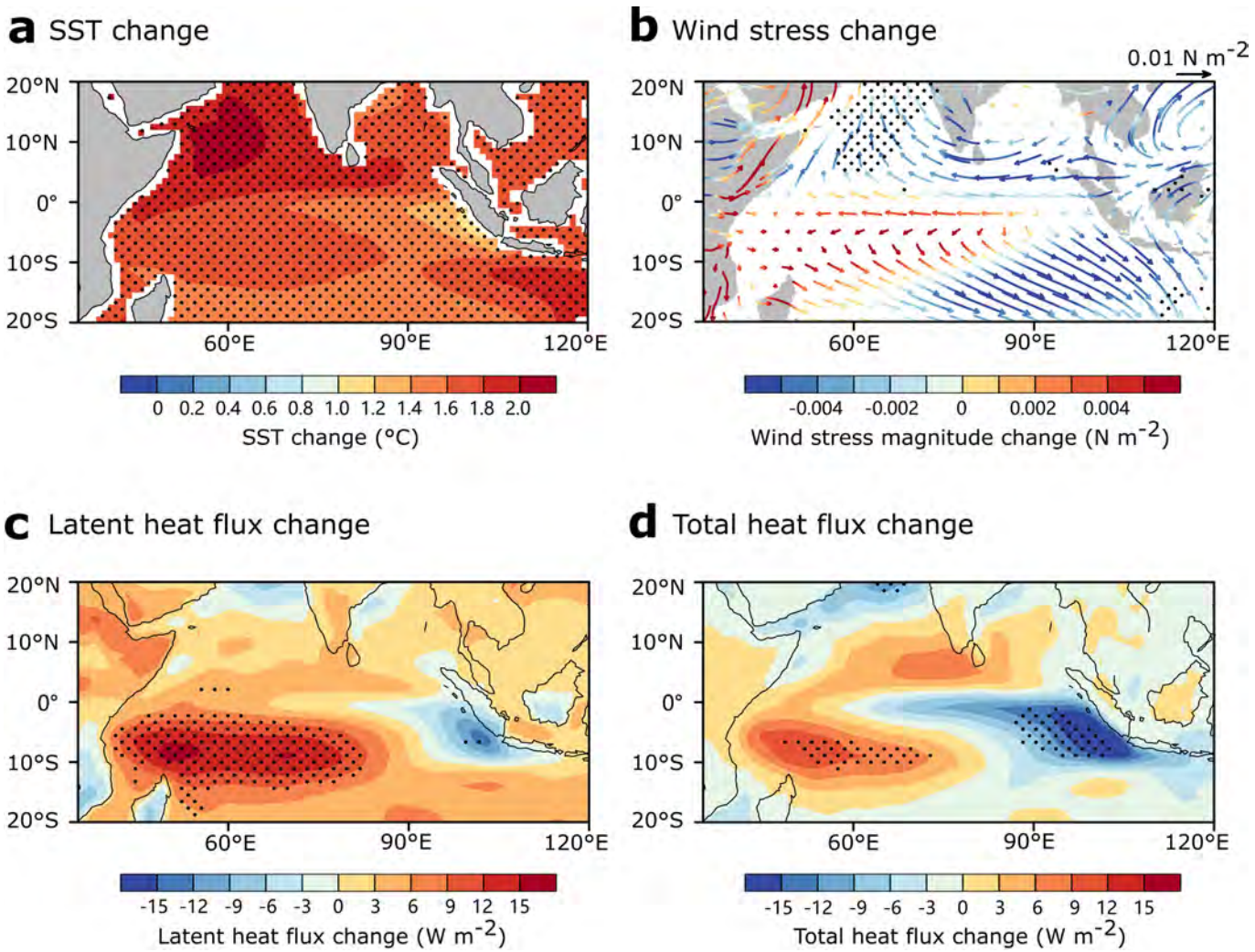

**Extended Data Fig. 7 | Sensitivity of projected changes in pIOD variability to model selections. a**, **b**, Comparison of variability of the moderate pIOD (M-index) and strong pIOD (S-index), respectively, in the present-day (1900–1999) and future (2000–2099) climate in 30 models that simulate a nonlinear coefficient $\alpha$ greater than 33.3% of the observed value. In **a**, a total of 27 out of the 30 models (90%) simulate a decrease in M-index variability in the future period (red bars) from that in the present-day period (blue bars). This leads to a multi-model mean decrease of 17%, statistically significant above the 95% confidence level, as indicated by the error bars. In **b**, a total of 23 out of the 30 selected models (77%) simulate an increase in S-index variability in the future climate, with a multi-model mean increase of 19%, statistically significant above the 95% confidence level. Models from CMIP6 are indicated in purple. Models that simulate an opposite change to the multi-model mean are indicated by green circles.




**Extended Data Fig. 8 | Sensitivity of projected changes in pIOD variability to emission scenario.** Shown are results for emission scenario RCP 4.5. **a**, **b**, Comparison of variability of the moderate pIOD (M-index) and strong pIOD (S-index), respectively, in the present-day (1900–1999) and future (2000–2099) climate in 15 out of 33 CMIP5 models that simulate a nonlinear coefficient $\alpha$ greater than 50% of the observed value. In **a**, a total of 14 out of the 15 models (93%) simulate a decrease in M-index variability in the future period (red bars) from that in the present-day period (blue bars). This leads to a multi-model mean decrease of 13%, statistically significant above the 95% confidence level, as indicated by the error bars. In **b**, a total of 13 out of the 15 selected models (87%) simulate an increase in S-index variability in the future climate, with a multi-model mean increase of 16%, statistically significant above the 95% confidence level. Models that simulate an opposite change to the multi-model mean are indicated by green circles. A total of 33 CMIP5 models are used here as outputs from CMCC-CESM and FGOALS-g2 are unavailable.




**Extended Data Fig. 9 | Relationship between projected changes in ENSO and in pIOD. a**, Scatter plot of changes in September, October, and November (SON) S-index variability versus changes in SON Niño3.4 index variability. The change is defined as the difference between the present-day (1900–1999) and future (2000–2099) period scaled by the corresponding increase in global mean SST in each model. **b**, Same as **a**, but between changes in SON M-index variability and in SON Niño3.4 index variability. **c, d**, Same as **a, b**, but for changes in strong pIOD (S-index>1.5 s.d.) or moderate pIOD (M-index>1.25 s.d.) frequency (events per 100 years) vs frequency of El Niño (Niño3.4>1.0 s.d.). The associated correlation coefficient and *P*-value are plotted. There is no relationship between changes in El Niño and S-index in **a** and **c**. Although there is a positive correlation in **b** and **d**, the changes in M-index and in El Niño are in the opposite direction, contravening the relationship between El Niño and M-index.




**Extended Data Fig. 10 | Projected mean state change in the tropical Indian Ocean.** Shown are averages over the 20 models that simulate a nonlinear coefficient $\alpha$ greater than 50% of the observed value. **a**, Multi-model ensemble averaged mean state changes of SON SST (in °C) between the future (2000–2099) and the present-day (1900–1999) climate. **b**, Same as **a**, but for the mean state change in surface wind stress (in N m$^{-2}$); changes in wind stress magnitude (in N m$^{-2}$) are indicated by colors. **c**, Same as **a**, but for the mean state change in latent heat flux (in W m$^{-2}$). **d**, Same as **a**, but for the mean state change in total heat flux (in W m$^{-2}$); positive values indicate an upward transfer of energy. Stippled areas indicate where changes are statistically significant above the 95% confidence level according to a two-tailed Student's *t*-tests.

OCEANOGRAPHY

# Dependence of regional ocean heat uptake on anthropogenic warming scenarios

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**The North Atlantic and Southern Ocean exhibit enhanced ocean heat uptake (OHU) during recent decades while their future OHU changes are subject to great uncertainty. Here, we show that regional OHU patterns in these two basins are highly dependent on the trajectories of aerosols and greenhouse gases (GHGs) in future scenarios. During the 21st century, North Atlantic and Southern Ocean OHU exhibit similarly positive trends under a business-as-usual scenario but respectively positive and negative trends under a mitigation scenario. The opposite centurial OHU trends in the Southern Ocean can be attributed partially to distinct GHG trajectories under the two scenarios while the common positive centurial OHU trends in the North Atlantic are mainly due to aerosol effects. Under both scenarios, projected decline of anthropogenic aerosols potentially induces a weakening of the Atlantic Meridional Overturning Circulation and a divergence of meridional oceanic heat transport, which leads to enhanced OHU in the subpolar North Atlantic.**

## INTRODUCTION

Since the industrial era, Earth's energy imbalance has largely been caused by changes in greenhouse gases (GHGs) and aerosols due to human activities (*1*). Global oceans absorb 93% of the net energy (*1*) through air-sea heat flux or in the form of the so-called ocean heat uptake (OHU). Regional OHU change is of great importance for regional sea level change (*2*), transient climate sensitivity and related feedbacks (*3*, *4*), and meridional displacement of the intertropical convergence zone (*5*, *6*). Throughout the historical period (1850–2005), the broad Southern Ocean has accounted for a majority of global OHU while the North Atlantic has taken a relatively smaller share (*7*–*9*). This asymmetry of OHU and ocean heat distribution could be attributed largely to the concurrent increases in GHGs and aerosols during the past century (*10*–*12*) and also to the alterations of deep ocean circulations in these two basins (*13*–*21*). In the North Atlantic, the warming effect of increasing GHGs leads to a weakening of the Atlantic Meridional Overturning Circulation (AMOC) and an increase of OHU, while the cooling effect of increasing aerosols has opposite influences on the AMOC and OHU (*22*, *23*). As a result of the compensation between the effects of aerosols and GHGs, the North Atlantic has shown a relatively small OHU increase during the historical period. On the other hand, anthropogenic aerosols are mainly confined to the Northern Hemisphere (*24*), which is quite different from GHGs that are homogeneously distributed over the globe. Over the Southern Ocean, increasing GHGs have served as one key factor in determining the intensive deep-water upwellings and enhanced OHU (*19*), whereas aerosols have had minimal impacts in this region.

Future emissions of GHGs and aerosols in the 21st century will be different from those in the historical period, as quantified by the Coupled Model Intercomparison Project Phase 5 (CMIP5) Representative Concentration Pathways (RCPs) (*25*). For example, the RCP8.5 scenario is a business-as-usual scenario while the RCP2.6 scenario is a mitigation scenario close to the target of the 2015 Paris Agreement that limits global warming of 1.5°C above preindustrial levels by 2100 (*26*). Both scenarios feature large and similar reductions in anthropogenic aerosol emissions (e.g., ~80% reduction in $SO_2$ emission by 2100) (*25*) but include different changes in GHG concentrations. For example, atmospheric $CO_2$ concentration will peak around 2050 and then decrease under the RCP2.6 scenario but will continue to increase throughout the 21st century under the RCP8.5 scenario (*25*). Such different evolutions of aerosols and GHGs will lead to a large scenario-driven uncertainty in regional OHU, indicating that future regional OHU change under the RCP2.6 scenario could be notably different from that under the RCP8.5 scenario (*12*, *22*, *27*). Nevertheless, the characteristics and underlying mechanisms of this scenario-driven uncertainty in regional OHU remain unclear.

## RESULTS

### Scenario-driven uncertainty in regional OHU

To explore the scenario-driven uncertainty in regional OHU, we examine CMIP5 RCP2.6 and RCP8.5 simulations. We use the multimodel ensemble mean (MME) of 26 CMIP5 models (table S1 and Materials and Methods) to minimize the effects of internal variability and intermodel uncertainty. We examine the long-term trends (Materials and Methods) of regional OHU during the 21st century under both scenarios (Fig. 1). Under the RCP2.6 scenario, OHU exhibits generally opposite trends between hemispheres: a significant positive trend in the subpolar North Atlantic but a negative trend in the Southern Ocean during 2006–2100 (Fig. 1A). These OHU trend patterns are consistent with the time series of regionally averaged OHU (Fig. 1C). In the subpolar North Atlantic, the averaged OHU increases rapidly before 2050 (0.139 W $m^{-2}$ $year^{-1}$, $P < 0.01$) and levels off or declines slightly after that, which corresponds to a positive trend of 0.045 W $m^{-2}$ $year^{-1}$ ($P < 0.01$) during 2006–2100. On the other hand, in the Southern Ocean, the averaged OHU shows a very weak increase by the late 2030s (0.035 W $m^{-2}$ $year^{-1}$, $P < 0.01$) but a large decrease afterward, which corresponds to a negative trend of −0.017 W $m^{-2}$ $year^{-1}$ ($P < 0.01$) during 2006–2100. Note that Southern Ocean OHU begins to decrease one to two decades earlier

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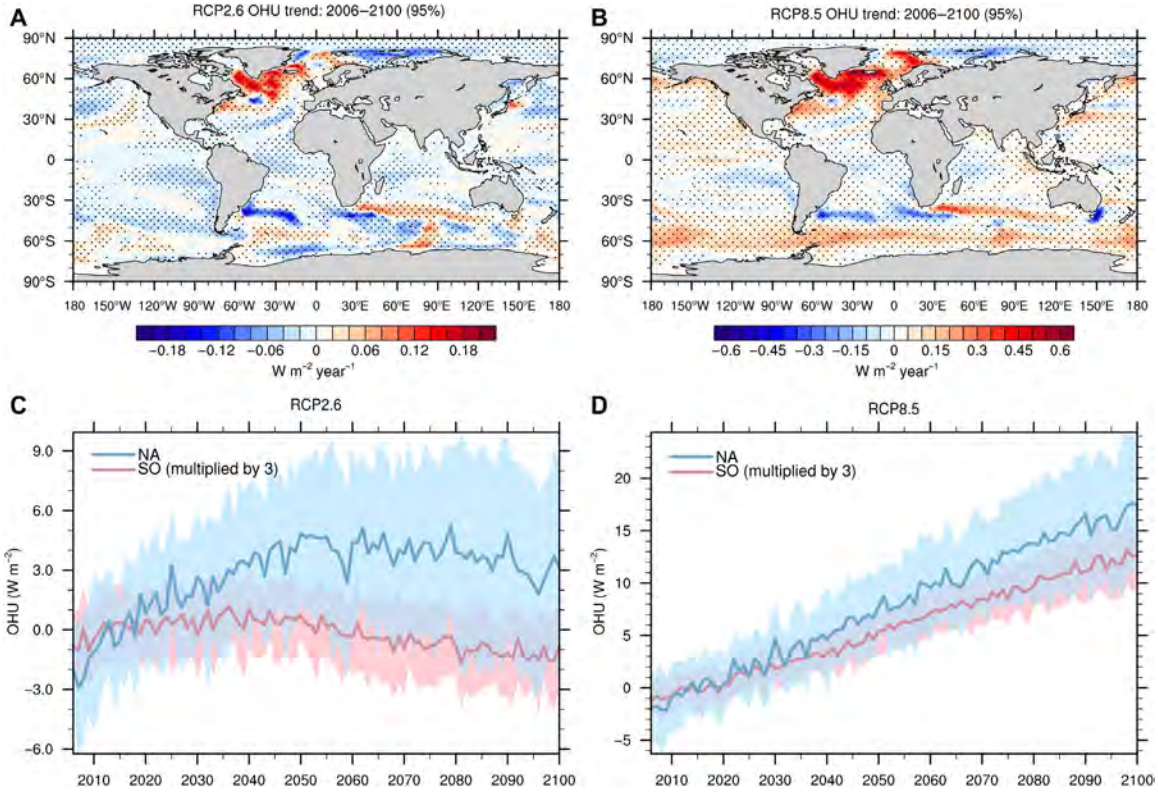



**Fig. 1. CMIP5-simulated OHU changes under the RCP2.6 and RCP8.5 scenarios.** Top row: CMIP5 MME OHU trends during 2006–2100 under the (**A**) RCP2.6 and (**B**) RCP8.5 scenarios. Stippling denotes where the OHU trends are statistically significant at the 95% confidence level based on the Student's *t* test. Positive (negative) values mean ocean gains (losses) heat. Bottom row: Annual mean OHU anomalies of CMIP5 model averaged over the North Atlantic (35°N to 70°N, 80°W to 10°W; MME, blue curve; model spread, blue shading) and averaged over the Southern Ocean (35°S to 70°S, 0° to 360°; MME, pink curve; model spread, pink shading) during 2006–2100 under the (**C**) RCP2.6 and (**D**) RCP8.5 scenarios. Anomalies are relative to the 2006–2025 mean. Model spread is represented by 1 SD of model results. Southern Ocean OHU anomalies are multiplied by a factor of 3 for a better illustration.

than the decline of $CO_2$ as projected under the RCP2.6 scenario, which might be associated with the stratospheric ozone recovery since the 2000s (*28*). In contrast, under the RCP8.5 scenario, OHU in both basins shows significant positive trends during 2006–2100, with the stronger one in the North Atlantic (Fig. 1B). For both basins, the regionally averaged OHU increases rapidly throughout the current century, with a larger trend in the North Atlantic (0.209 W $m^{-2}$ $year^{-1}$, $P < 0.01$) relative to the Southern Ocean (0.154 W $m^{-2}$ $year^{-1}$, $P < 0.01$) (Fig. 1D). Here, it merits attention that time series of regionally averaged OHU are not strictly linear in both the North Atlantic and Southern Ocean under both the RCP2.6 and RCP8.5 scenarios, which are largely due to complex trajectories of aerosols and GHGs. To explore the scenario-driven uncertainty in regional OHU, we resort to a common metric—the OHU trend during 2006–2100—to compare the primary regional OHU changes on a centurial time scale between the two scenarios. To summarize, the CMIP5 MME results suggest that regional OHU projections vary substantially between the mitigation and business-as-usual scenarios, which is potentially owing to the different trajectories of anthropogenic aerosols and GHGs between the two scenarios.

In addition, we investigate the relationship between aerosol radiative effect and regional OHU in CMIP5 models. In each model, the total aerosol radiative effect is estimated as aerosol effective radiative forcing (ERF) (Materials and Methods). We demonstrate the relationship between aerosol ERF and regional OHU trend during the 21st century among 15 of the 26 CMIP5 models according to the availability of model aerosol ERF (table S1) (*6*). Over the Southern Ocean, the correlation between aerosol ERF and OHU trend is not significant under both scenarios (Fig. 2, C and D), indicative of a potentially small influence of aerosols on the Southern Ocean OHU, while in the North Atlantic, a significant positive correlation between aerosol ERF and OHU trend is evident under either the RCP2.6 or the RCP8.5 scenario (Fig. 2, A and B), which means that models with stronger aerosol ERFs generally simulate larger centurial OHU trends in the North Atlantic. In addition, the correlation coefficient is larger under the RCP2.6 scenario than that under the RCP8.5 scenario. Under the RCP2.6 scenario, models with no aerosol indirect effects (Materials and Methods) show the smallest OHU trends, while models with both aerosol indirect effects (Materials and Methods) show the largest OHU trends (Fig. 2A). As such, our result confirms the importance of aerosol effects on the North Atlantic OHU, especially for the RCP2.6 scenario, although it might be partially affected by the implicit tuning of models over the historical period due to uncertainty in aerosol ERF and climate sensitivity (*29*, *30*).

## Roles of anthropogenic aerosols and GHGs in modulating regional OHU

To further explore the roles of anthropogenic aerosols and GHGs in modulating regional OHU under the RCP2.6 and RCP8.5 scenarios, we choose a broadly used CMIP5 model—the Community Earth

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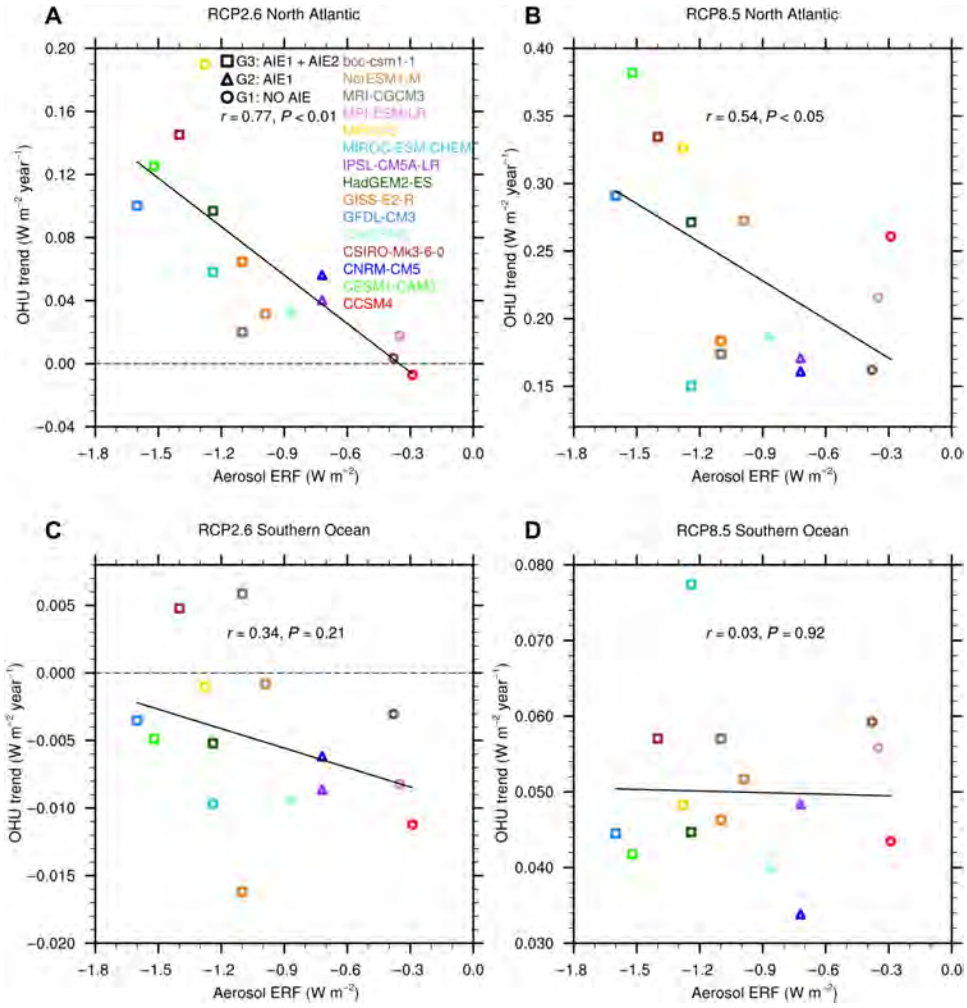



**Fig. 2. Relationships between aerosol ERFs and regional OHU trends in CMIP5 models over the 21st century under the RCP2.6 and RCP8.5 scenarios.** Top row: Aerosol ERFs of 2000 relative to 1850 and North Atlantic OHU trends of 2006–2100 under the (**A**) RCP2.6 and (**B**) RCP8.5 scenarios in CMIP5 models. Bottom row: As in the top row but for aerosol ERFs and Southern Ocean OHU trends (**C** and **D**). Models are denoted by different colors. Models with no aerosol indirect effects (G1: NO AIE) are denoted by circles. Models with only the first aerosol indirect effects, cloud albedo effects (G2: AIE1), are denoted by triangles. Models with both aerosol indirect effects, cloud albedo and cloud lifetime effects (G3: AIE1 + AIE2), are denoted by squares. Solid black line in each panel represents the linear regression line for the scatters. Correlation coefficient (*r*) between ERFs and OHU trends and *P* value are noted in each case. Notice the different scales in each panel.

System Model version 1 with the Community Atmosphere Model version 5 (CESM1-CAM5; Materials and Methods). CESM1-CAM5 includes both aerosol indirect effects and shows an aerosol ERF of −1.52 W $m^{-2}$ in 2000 relative to 1850, which is strong (table S1) but still within the range of uncertainty estimates of CMIP5 models (for CMIP5 models, 90% confidence level of aerosol ERF corresponds to −2.0 to −0.4 W $m^{-2}$) (*31*). Given the importance of aerosols to North Atlantic OHU, CESM1-CAM5 projects a large OHU change in the North Atlantic at least partially due to its strong aerosol ERF (Fig. 2, A and B). Nevertheless, this model does well capture the characteristics of centennial OHU evolution patterns as in CMIP5 MME for both North Atlantic and Southern Ocean under both the RCP2.6 and RCP8.5 scenarios. Using CESM1-CAM5, we conduct parallel sensitivity experiments with fixed present-day anthropogenic aerosol emissions (rcpFA26 and rcpFA85 hereafter; Materials and Methods) based on the archived all-forcing CESM1-CAM5 simulations for the RCP2.6 and RCP8.5 scenarios (rcp26 and rcp85 hereafter; Materials and Methods). The difference between the two experiments in each pair (rcp26-rcpFA26 or rcp85-rcpFA85) then yields the net climate effect of future anthropogenic aerosols reduction under each scenario.

We find that major differences of regional OHU changes between rcp26 and rcpFA26 occur in the North Atlantic. OHU exhibits a much stronger positive centurial trend in the subpolar North Atlantic in rcp26 than that in rcpFA26 (Fig. 3, A and C). The regionally averaged OHU time series show that North Atlantic OHU in rcp26 increases throughout the century (0.126 W $m^{-2}$ $year^{-1}$, $P < 0.01$) with a larger rate before 2050 (0.243 W $m^{-2}$ $year^{-1}$, $P < 0.01$), while North Atlantic OHU in rcpFA26 increases before the 2040s and then declines until 2100 (Fig. 4A). On the other hand, OHU changes in the Southern Ocean are highly similar between the two simulations. In both rcp26 and rcpFA26, Southern Ocean OHU exhibits generally negative centurial trends during 2006–2100 (Fig. 3, A and C). The regionally averaged OHU time series depict that Southern Ocean OHU peaks around the 2030s and diminishes afterward (Fig. 4C).

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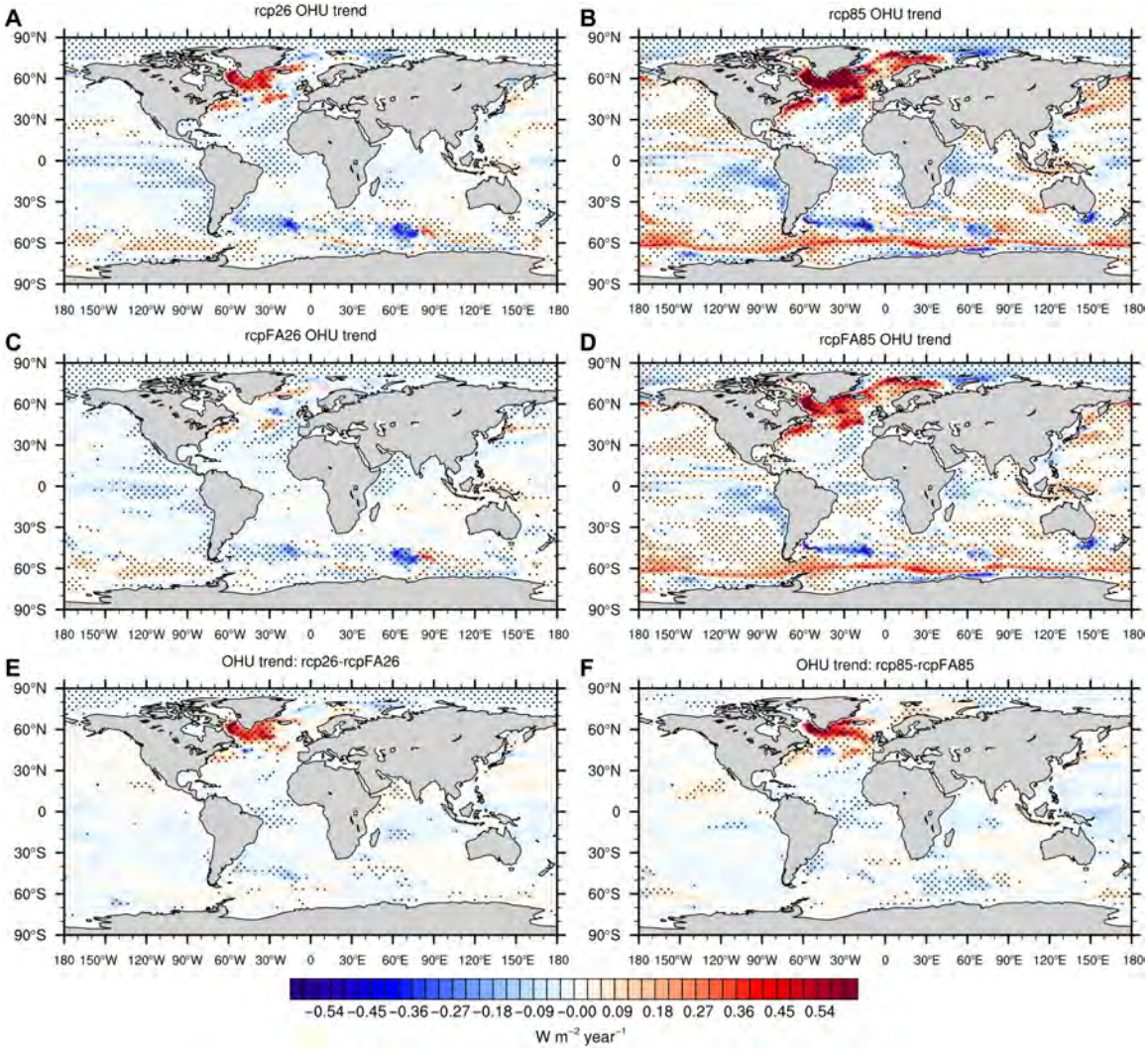



**Fig. 3. CESM1-CAM5 OHU trends under the RCP2.6 and RCP8.5 scenarios and aerosol effects on OHU trends.** Left column: OHU trends during 2006–2100 (**A**) in rcp26 and (**C**) rcpFA26 simulations with CESM1-CAM5 and (**E**) for their difference (rcp26-rcpFA26), which indicates aerosol effects. Right column: As in the left column but for (**B**) rcp85 and (**D**) rcpFA85 and (**F**) their difference. Stippling denotes where trends are statistically significant at the 95% confidence level based on the Student's *t* test. Positive (negative) values mean ocean gains (losses) heat.

The comparison between rcp26 and rcpFA26 suggests that the interhemispheric contrast of future OHU change is evident only when aerosol effects are included. This is because anthropogenic aerosols mainly concentrate in the Northern Hemisphere and are projected to decline under the RCP2.6 scenario (figs. S1 and S2; see Supplementary Materials), which produces an extra warming effect superimposed on the GHG effect. In the North Atlantic, OHU response is primarily attributed to the total effects of anthropogenic aerosols and GHGs. In rcp26, North Atlantic OHU increases at a larger rate before 2050 and then continues to increase until 2100 (Fig. 4A). This is consistent with the extra aerosol warming effect under the RCP2.6 scenario, as reflected by the aerosol evolution of large decrease before 2050 and continued reduction until 2100 (fig. S2). However, in rcpFA26, North Atlantic OHU decreases after 2050 (Fig. 4A), following the GHG evolution under the RCP2.6 scenario. Differences of North Atlantic OHU between rcp26 and rcpFA26 can be seen from the significant positive long-term trend during 2006–2100 within the subpolar North Atlantic in rcp26-rcpFA26 (Fig. 3E). Over the Southern Ocean, the OHU difference between rcp26 and rcpFA26 is not robust (Figs. 3E and 4C) primarily due to a minimal effect of anthropogenic aerosols.

We also compare rcp85 and rcpFA85 and find that the significant OHU differences due to aerosol effects again occur in the subpolar North Atlantic (Fig. 3F). As anthropogenic aerosols decline under the RCP8.5 scenario (fig. S2), North Atlantic OHU in rcpFA85 shows a smaller positive trend than that in rcp85 (Fig. 3, B and D), largely owing to the lack of aerosol-induced warming. Meanwhile, this difference can be seen from regional averaged OHU time series in which North Atlantic OHU increases at a smaller rate in rcpFA85 (0.241 W $m^{-2}$ $year^{-1}$, $P < 0.01$) than rcp85 (0.379 W $m^{-2}$ $year^{-1}$, $P < 0.01$) during 2006–2100 (Fig. 4B). Over the Southern Ocean, anthropogenic aerosol effects are minimal, while GHG effects are major. Southern Ocean OHU exhibits highly similar evolution patterns in both rcpFA85 and rcp85, featuring continuous increases during

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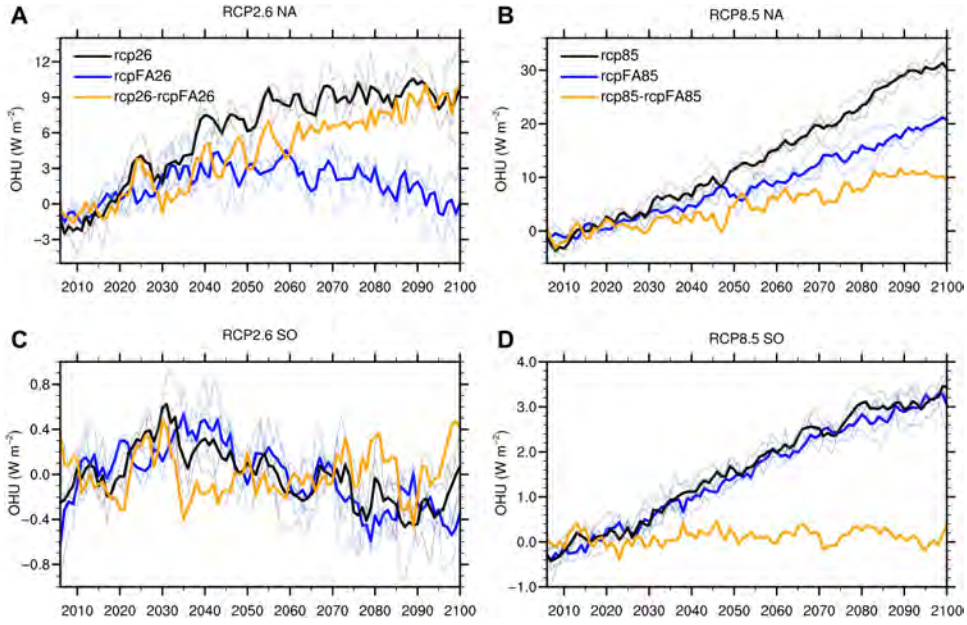



**Fig. 4. CESM1-CAM5 regional OHU time series under the RCP2.6 and RCP8.5 scenarios.** Top row: Annual mean OHU anomalies averaged over the North Atlantic (NA) (35°N to 70°N, 80°W to 10°W) during 2006–2100 under the (**A**) RCP2.6 and (**B**) RCP8.5 scenarios. Bottom row: As in the top row but for annual mean OHU anomalies averaged over the Southern Ocean (SO) (35°S to 70°S, 0° to 360°) (**C** and **D**). Thick black and blue curves indicate OHU anomalies in rcp and rcpFA simulations with CESM1-CAM5, which are calculated from the ensemble means of individual simulations. Thin black and blue curves denote OHU anomalies for three ensemble members in each of the rcp and rcpFA simulations. Orange lines denote the difference between the ensemble means of rcp and rcpFA simulations, which indicate aerosol effects. OHU anomalies are relative to the 2006–2025 mean and a 5-year running mean is applied. Notice the different scales in each panel.

2006–2100 (Fig. 4D; 0.041 and 0.042 W m$^{-2}$ year$^{-1}$, respectively, both $P < 0.01$). To summarize, our results from CESM1-CAM5 experiments suggest that aerosol effects play an important role in the North Atlantic in future OHU changes under both the RCP2.6 and RCP8.5 scenarios. Projected decline of anthropogenic aerosols enhances OHU in the subpolar North Atlantic, resulting in positive long-term trends during 2006–2100 under both scenarios. Over the Southern Ocean, the effect of aerosols is minimal, while the effect of GHGs is major such that the projected OHU shows negative and positive long-term trends during 2006–2100 under the RCP2.6 and RCP8.5 scenarios, respectively.

## Effect of aerosols on ocean heat storage via altering ocean circulation

We discover that the effect of aerosols not only acts on the OHU at Atlantic Ocean surface but also extends deep into the interior ocean and affects ocean heat storage (OHS) through altering ocean circulation and heat transport across the basin. Since 2006, the reduced anthropogenic aerosols trigger a weakened AMOC under both the RCP2.6 and RCP8.5 scenarios (fig. S3), which likely follows the mechanism as described in (*13*). Although the anthropogenic aerosol forcing turns steady after 2050 under the RCP2.6 scenario (fig. S2), the weakening of the AMOC continues, primarily owing to the subsequent ocean adjustments in the interior Atlantic Ocean (*32*). Here, we examine the difference between rcp26 and rcpFA26 simulations during three subperiods: 2006–2050, 2051–2075, and 2076–2100. When comparing the two post-2050 periods (2051–2075 and 2076–2100), we find that the continued AMOC slowdown induces a further cooling of sea surface temperatures (SSTs) in the subpolar North Atlantic (figs. S4B and S5, E and I). The enhanced SST cooling causes a further increase of North Atlantic OHU (Fig. 4A and fig. S4B) primarily via a turbulent heat flux feedback (*33*, *34*) and, meanwhile, a further reduction of evaporation to the south of Greenland (fig. S6, E and H). As a dominant factor, the intensified reduction of evaporation enlarges the surface freshwater input (fig. S6, D and G), acting to freshen the surface water to the south of Greenland. Although either surface cooling (fig. S5, E and I) or freshening (fig. S5, F and J) is present in this deep convection region, the haline effect dominates, which leads to a reduction of local sea surface density (fig. S5, G and K), a suppression of deep convection (fig. S5, H and L), and hence a further weakening of the AMOC (fig. S4B). As this loop continues, SST cooling and OHU increase persist in the subpolar North Atlantic after 2050. Under the RCP8.5 scenario, similar processes also operate (figs. S4, S7, and S8).

We further conduct an ocean heat budget analysis on the Atlantic (*18*, *20*, *21*, *35*, *36*) during 2006–2100 for either scenario using CESM1-CAM5 simulations, which incorporates OHU, OHS, meridional ocean heat transport (OHT), and diffusive processes into a dynamically consistent framework (Materials and Methods). From the difference of the 2006–2100 averages of rcp26 and rcpFA26, we find that the aerosol-driven AMOC change shows the maximum weakening at ~40°N, around the latitude of climatological AMOC maximum under the RCP2.6 scenario (Fig. 5E and fig. S3). This AMOC decline brings about a reduced northward OHT across the Atlantic primarily through its Eulerian-mean component (Fig. 5G). The magnitude of OHT reduction markedly diminishes poleward from ~40°N, leading to a meridional divergence of OHT in the subpolar North Atlantic (Fig. 5I). The OHT divergence acts to cool the whole column ocean waters in the subpolar North Atlantic (fig. S9A), which triggers more heat uptake from the atmosphere via ocean surface (Fig. 5, A and I) to

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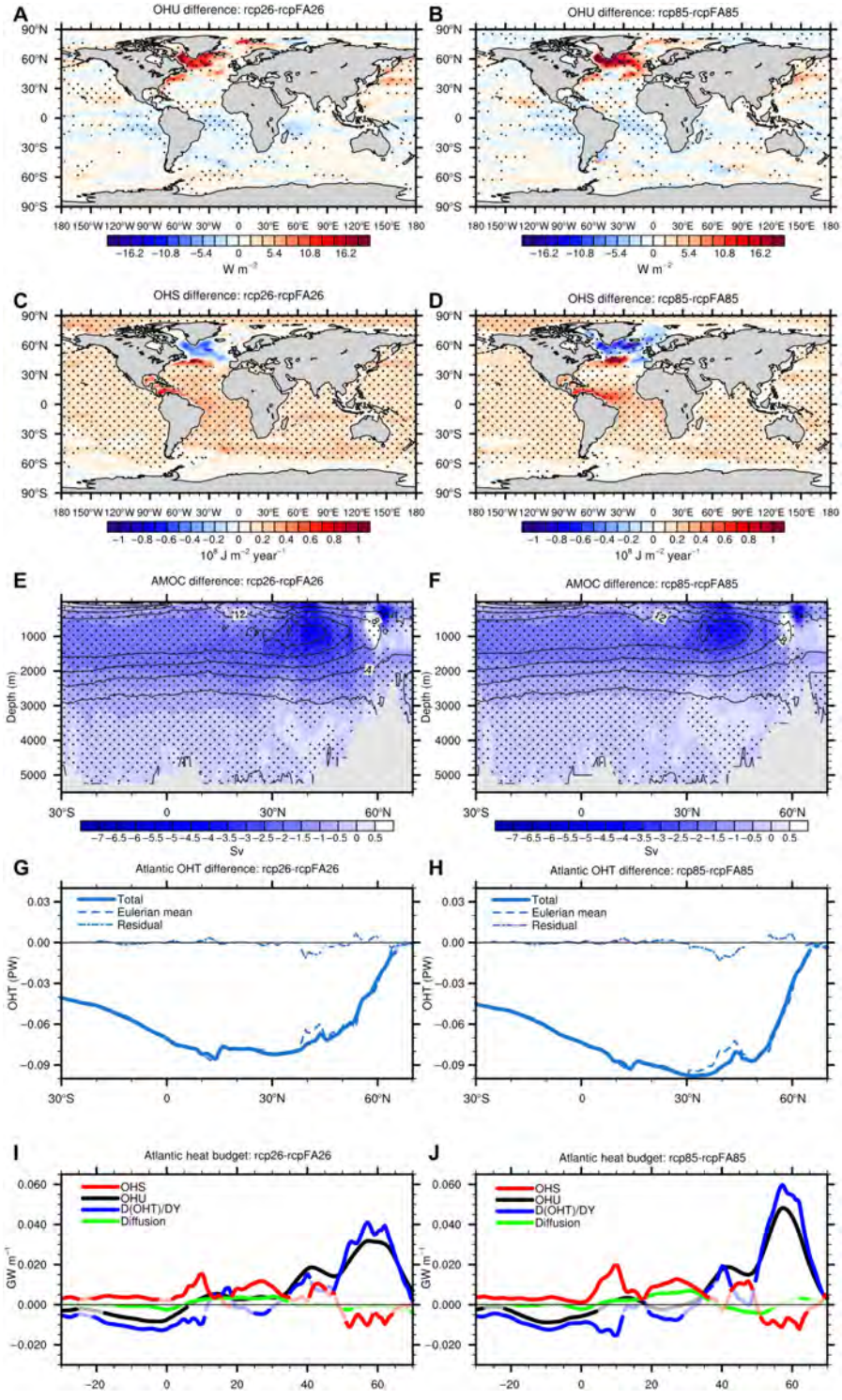



**Fig. 5. CESM1-CAM5 aerosol effects on OHU and ocean heat redistribution under the RCP2.6 (left) and RCP8.5 (right) scenarios.** Left column: Differences of annual mean (**A**) OHU (positive into the ocean); (**C**) OHS (positive values mean increases of heat in the ocean); (**E**) meridional overturning streamfunction in the Atlantic (negative values denote the weakening of meridional circulation); (**G**) OHT in the Atlantic (positive values denote anomalous northward heat transport): heat transport induced by total advection (solid), Eulerian-mean advection (dashed), and their residual (dotted; including eddy-induced advection and sub-mesoscale advection); (**I**) heat budget in the Atlantic: zonally integrated full-depth OHS (red), zonally integrated OHU (black), divergence of OHT induced by advections (blue), and zonally integrated full-depth heat change due to diffusive processes (green) during 2006–2100 under the RCP2.6 scenario. Right column: As in the left column but for the RCP8.5 scenario (**B**, **D**, **F**, **H**, and **J**). Stippling in the top six panels denotes where differences are statistically significant at the 95% confidence level based on the Student's *t* test. In the bottom four panels, dark colors (relative to light colors) denote where differences are statistically significant at the 95% confidence level based on the Student's *t* test. Contours in (E) and (F) denote the annual mean AMOC averaged over 2006–2100 under the RCP2.6 and RCP8.5 scenarios, respectively. Curves in (I) and (J) are derived after a 7.5° running mean.

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compensate for this dynamically induced cooling (*37*). Nevertheless, the compensation from enhanced OHU is incomplete, and meanwhile, a small amount of heat can escape from the subpolar region through diffusive processes. As a result, the storage of oceanic heat diminishes in the subpolar North Atlantic (Fig. 5, C and I). Here, note that in the subpolar North Atlantic, the diminished OHS represents a cooling tendency of full-depth ocean waters (fig. S9A) as consistent with the anomalous cooling of SSTs (fig. S10E).

Under the RCP8.5 scenario, the aerosol effect is similar to that under the RCP2.6 scenario. Enhanced OHU is evident over the subpolar North Atlantic (Fig. 5, B and J). The reduction of anthropogenic aerosols induces a weaker and shallower AMOC (Fig. 5F and fig. S3) and results in a redistribution of oceanic heat within the Atlantic. In particular, heat is moved from high-latitude ocean to low-latitude ocean such that OHS decreases in the former region but increases in the latter (Fig. 5D). The significantly decreased OHS in the subpolar North Atlantic is tied to the strong cooling of local full-depth ocean waters (fig. S9B), which is consistent with the cooling of local SSTs (fig. S10F).

## DISCUSSION

Our study suggests that during the 21st century, the projected OHU shows respectively positive and negative trends in the North Atlantic and Southern Ocean under the RCP2.6 scenario but collectively positive trends within the two basins under the RCP8.5 scenario. Such large scenario-driven uncertainty in regional OHU projections is tightly related to different evolution patterns of anthropogenic aerosols and GHGs under both scenarios. In the RCP2.6 scenario, anthropogenic aerosol reduction potentially serves as a main factor contributing to the enhanced OHU in the North Atlantic, which is evident from the significant positive correlation between aerosol ERF and OHU trend among CMIP5 models. Over the Southern Ocean, the decreased OHU after the 2030s could be related to the projected GHG decline in the latter half of the century and stratospheric ozone recovery since the 2000s. In the RCP8.5 scenario, both anthropogenic aerosol reduction and GHG increase contribute to the OHU increase in the North Atlantic. Over the Southern Ocean, the rising GHGs promote the OHU increase throughout the 21st century given that the strong GHG effect in this scenario likely overwhelms the effect of ozone recovery. Under both the RCP2.6 and RCP8.5 scenarios, aerosol effects induce an enhanced OHU but diminished storage of oceanic heat in the subpolar North Atlantic. This is due to an aerosol-induced AMOC slowdown, which engenders a redistribution of oceanic heat from high latitudes to low latitudes.

Besides aerosol changes, other forcings such as GHG changes can also induce hemispheric contrast of SSTs and hence OHU. One example is the SST warming minimum or the so-called North Atlantic warming hole (*38*) developed to the south of Greenland under increasing $CO_2$ (*3*, *39*), which is suggested to be associated with an AMOC slowdown. A similar feature can be found under the RCP8.5 scenario when GHG forcing is exclusively considered (fig. S10D). This GHG-driven warming minimum, however, exhibits a smaller magnitude than that in the full SST response to the total RCP8.5 forcing. This is because the strong SST cooling driven by future anthropogenic aerosol reduction reinforces the warming minimum to the south of Greenland, making the warming hole more evident in the full SST response (fig. S10). Moreover, beyond the RCPs used in CMIP5, multiple shared socioeconomic pathway scenarios have been introduced to current CMIP6 models, which span much broader aerosol emission pathways (*40*). Improved understanding of how anthropogenic emissions affect OHU and ocean circulations under different scenarios is necessary to reduce the uncertainty of climate projections.

## MATERIALS AND METHODS

### Datasets

Model results from the CMIP5 of the Intergovernmental Panel on Climate Change (*41*) are used to examine OHU trends and its evolution in the North Atlantic and Southern Ocean over 2006–2100 (Fig. 1). Here, based on the availability of surface heat fluxes (SHFs) over 2006–2100, we use 26 CMIP5 models, in which 62 members of CMIP5 RCP2.6 simulation and 66 members of CMIP5 RCP8.5 simulation are analyzed (table S1). $CO_2$ concentrations in the RCP2.6 and RCP8.5 scenarios (*42*, *43*) are available from the RCP database website (http://tntcat.iiasa.ac.at:8787/RcpDb/dsd?Action=htmlpage&page=about).

### Analysis methods

The CMIP5 MME provides an estimate of climate response to external forcing, as internal climate variability is averaged out. For each scenario, we first calculate the ensemble member mean for each model and then average the mean values from all the models to achieve the MME of CMIP5 models. This approach allows equal weighting for each model in the calculation of MME.

The least-square linear regression is used at each grid box to estimate the long-term trends over 2006–2100 for future scenarios. Long-term trends of the difference between "rcp" and "rcpFA" simulations for OHU (Fig. 3), aerosol optical depth (fig. S1), zonal-mean Atlantic Ocean temperature (fig. S9), and SSTs (fig. S10) are calculated to reveal the net anthropogenic aerosol effect. The statistical significance of trends is performed with the Student's *t* test. In Fig. 5, we take the difference between the 2006–2100 means in rcp and rcpFA simulations at each grid box to obtain the net anthropogenic aerosol effect and then calculate the significance for the difference based on the Student's *t* test.

### Aerosol radiative effects

Aerosol radiative effects on climate include direct effects in which aerosols scatter and absorb shortwave radiation and indirect effects in which aerosols alter the cloud albedo (first) and cloud lifetime (second). The model representations of aerosol effects are used to further subdivide the CMIP5 models into three groups: those without aerosol indirect effects (NO AIE), those with only the first aerosol indirect effect (AIE1), and those with both aerosol indirect effects (AIE1 + AIE2) (table S1). For climate models, ERF is calculated from the top-of-the-atmosphere (TOA) flux differences between atmosphere-only simulations with identical SSTs but different atmospheric compositions (*44*), and aerosol ERF is calculated as the TOA flux difference between the atmosphere-only simulation with aerosol forcing and that without aerosol forcing. Models that include aerosol indirect effects generally have larger aerosol ERF than models that do not include aerosol indirect effects, and models with both aerosol indirect effects generally have larger aerosol ERF than those with only one aerosol indirect effect (Fig. 2).

### The NCAR CESM1-CAM5 and the fixed-aerosol experiments

The CESM1-CAM5 coupled climate model (*45*) from the National Center for Atmospheric Research (NCAR) is used to investigate the

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responses of OHU and ocean heat redistribution to the projected decline of anthropogenic aerosols under the RCP2.6 and RCP8.5 scenarios. The atmosphere component is the CAM5. The ocean component is the Parallel Ocean Program version 2. The land component is the Community Land Model version 4. The sea ice component is the Community Ice Code version 4. We use the f09_g16 version, with ~1° horizontal resolution and 30 vertical levels in the atmosphere component and ~1° horizontal resolution and 60 vertical levels in the ocean component. In CESM1-CAM5, the total surface freshwater flux into the ocean is computed as a sum of precipitation ($P$) minus evaporation ($-E$), river runoff ($R$), freshwater fluxes due to sea ice melting ($M$), and brine rejection ($Br$) and other sources.

We use the archived rcp simulations (rcp26 and rcp85) with CESM1-CAM5. Each rcp simulation is composed of three ensemble members over 2006–2100 and is available from the CMIP5 archives. Using CESM1-CAM5, we perform the rcpFA experiments (rcpFA26 and rcpFA85) from 2006 to 2100 under the same RCP scenarios except that the emissions of anthropogenic aerosols and their precursors (including aerosols from biomass burning) are fixed at 2005 levels with a constant annual cycle. All the other forcings are identical as in rcp simulations. For both rcpFA26 and rcpFA85 simulations, we conduct three-member ensemble simulations. Each member starts from 2006 with the initial condition obtained from the archived CESM historical simulations. Unless otherwise specified, the results of CESM simulations presented in this study are based on the three-member ensemble means of individual simulations. In model simulations, the AMOC index is defined as the maximum of streamfunction below 500 m in the Atlantic.

### Ocean heat budget analysis

OHU is defined as net downward SHF into the ocean, which is equal to the sum of net surface shortwave radiation flux, net surface longwave radiation flux, latent heat flux, sensible heat flux, and fluxes from other sources. OHS is defined as the tendency of ocean heat content (OHC) and denotes the rate of oceanic heat change for full-depth water column. OHT is defined as the transport of heat by advective processes, including Eulerian-mean, eddy-induced, and sub-mesoscale advective processes.

Along each latitude of the Atlantic, the full-depth ocean heat budget follows

$$\frac{\partial}{\partial t}\int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p\theta dz\,dx = \int_{X_w}^{X_E}(\mathrm{SHF})\,dx - \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p\ \nabla\cdot(\mathbf{v}\theta)\,dz\,dx + \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p D\,dz\,dx \quad (1)$$

$\rho_0 = 1026$ kg m$^{-3}$ is seawater density, $c_p = 3996$ J K$^{-1}$ kg$^{-1}$ is the specific heat capacity of seawater, $\theta$ is the potential temperature of seawater, $H$ is the depth of ocean bottom, and $X_W$ and $X_E$ denote the longitudes of the western and eastern boundaries of the Atlantic, respectively. $\nabla$ denotes the three-dimensional gradient operator and $\mathbf{v} = \overline{\mathbf{v}} + \mathbf{v}^*$, where $\overline{\mathbf{v}}$ is Eulerian-mean velocity and $\mathbf{v}^*$ denotes the sum of eddy-induced and sub-mesoscale velocities. $D$ denotes diffusive processes.

We then define OHS as

$$\mathrm{OHS} = \frac{\partial}{\partial t}\mathrm{OHC} = \frac{\partial}{\partial t}\int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p\theta\,dz\,dx \quad (2)$$

OHU as

$$\mathrm{OHU} = \int_{X_w}^{X_E}(\mathrm{SHF})\,dx \quad (3)$$

OHT as

$$\mathrm{OHT} = \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p(\overline{\nabla}\theta + \mathrm{v}^*\theta)\,dz\,dx = \overline{\mathrm{OHT}} + \mathrm{OHT}^* \quad (4)$$

and integrated heat change due to diffusive processes as

$$\mathrm{Diff} = \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p D\,dz\,dx \quad (5)$$

where $\overline{\mathrm{OHT}} = \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p\overline{\nabla}\theta dz\,dx$ denotes the OHT by Eulerian-mean advection ($\overline{\nabla}$ is Eulerian-mean meridional velocity) and $\mathrm{OHT}^* = \int_{X_w}^{X_E}\int_{-H}^{0}\rho_0 c_p\mathrm{v}^*\theta dz\,dx$ denotes the OHT by eddy-induced and sub-mesoscale advection (v* is the sum of eddy-induced and sub-mesoscale meridional velocities). Hence, the heat budget (Eq. 1) can be written as

$$\mathrm{OHS} = \mathrm{OHU} - \frac{\partial}{\partial y}\mathrm{OHT} + \mathrm{Diff} \quad (6)$$

As can be seen from Eq. 6, OHS is determined by OHU, meridional gradient (divergence) of OHT, and heat change due to diffusive processes.

## SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/6/45/eabc0303/DC1

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*Annual Review of Earth and Planetary Sciences*

# Responses of the Tropical Atmospheric Circulation to Climate Change and Connection to the Hydrological Cycle

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## Keywords

climate change, tropical atmospheric circulation, ocean precipitation, sea surface temperature, land rainfall, Hadley circulation

## Abstract

This review describes the climate change–induced responses of the tropical atmospheric circulation and their impacts on the hydrological cycle. We depict the theoretically predicted changes and diagnose physical mechanisms

for observational and model-projected trends in large-scale and regional climate. The tropical circulation slows down with moisture and stratification changes, connecting to a poleward expansion of the Hadley cells and a shift of the intertropical convergence zone. Redistributions of regional precipitation consist of thermodynamic and dynamical components, including a strong offset between moisture increase and circulation weakening throughout the tropics. This allows other dynamical processes to dominate local circulation changes, such as a surface warming pattern effect over oceans and multiple mechanisms over land. To improve reliability in climate projections, more fundamental understandings of pattern formation, circulation change, and the balance of various processes redistributing land rainfall are suggested to be important.

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## 1. INTRODUCTION

Global warming is likely to alter the distribution of water resources (Held et al. 2005, Seager et al. 2007, Zhang et al. 2007, Hegerl et al. 2015), and this redistribution will create socioeconomic, environmental, and security challenges. This is particularly true in the tropics, where almost half of the world's population and more than 80% of terrestrial biodiversity reside, and where societies are often extremely vulnerable to rainfall variability and change. Changes in the hydrological cycle at regional scales are closely related to the tropospheric circulation, which is primarily generated by the uneven distribution of diabatic heating and cooling (e.g., latent heating in convergence zones and radiative cooling in subsidence zones), and partly affected by synoptic eddy–induced heat and moisture transports (Pierrehumbert & Roca 1998, Schneider et al. 2010). Anthropogenic climate change modifies the stratification and diabatic forcing, and the large-scale circulation has to adjust accordingly to restore thermodynamic balance (e.g., Rodwell & Hoskins 1996). In this review, we examine how these adjustments occur and their likely impacts on the hydrological cycle.

Previous research focused on the effects of the global-mean increase of sea surface temperature (SST) as the zero-order problem, in order to define constraints for climate projections. The most important large-scale response of the circulation is a weakening in the tropical troposphere. Radiative and thermodynamic relations were proposed to diagnostically explain this slowdown (Knutson & Manabe 1995, Allen & Ingram 2002, Stephens & Ellis 2008), and the mean advection of stratification change (MASC) was more recently suggested (Ma et al. 2012) to explain how the atmospheric warming dynamically weakens the tropical circulation. Changes in the lapse rate have also been found to be responsible for the robust poleward expansion of the subsiding branches of the Hadley circulation in observations (Hu & Fu 2007) and general circulation model simulations (Frierson et al. 2007, Lu et al. 2007, Johanson & Fu 2009). However, these arguments are insufficient to explain why the intensity change of the Hadley cells is not as robust as that of the Walker cell (Vecchi & Soden 2007b).

On very large averaged spatial scales, regional precipitation change has been interpreted as a wet-get-wetter, dry-get-drier pattern, with rainfall increases at the core of existing rainy regions and decreases in current dry areas (Held & Soden 2006) and at the convective margins (Neelin et al. 2003, Chou & Neelin 2004, Chou et al. 2009). However, the dry-get-drier argument has been questioned (Scheff & Frierson 2012), because reduced precipitation appears along the outer flanks of the subtropics due to the poleward expansion of the subtropical dry zones. Nor does the wet-get-wetter interpretation hold at the smaller country-level scales relevant to climate change impacts (Chadwick et al. 2013, Greve et al. 2014, Roderick et al. 2014), where rainfall changes are strongly associated with circulation change. At regional scales, the dominant changes are often shifts in the

**Sea surface temperature (SST):** the water temperature close to the ocean's surface, often measured at 1 m depth

**Hadley circulation:** a large-scale tropical circulation, with air rising near the equator and descending in the subtropics and flowing equatorward near the surface and poleward near the tropopause

positions of convective regions, with associated changes in rainfall and water availability. These shifts are associated with a number of mechanisms that differ between land and ocean.

The wet-get-wetter, dry-get-drier hypothesis implicitly assumes a spatially uniform SST increase (SUSI); however, spatial variations in surface ocean warming are of considerable magnitude (Xie et al. 2010), with coherent seasonal variability (Sobel & Camargo 2011). Substantial research attention has therefore been downscaled to regional climate change (Xie et al. 2015) as the first-order problem. A warmer-get-wetter paradigm has been recognized as essential for the tropical ocean, casting SST patterns—i.e., deviations from the tropical-mean SST increase—as an important parameter for the adjustment in atmospheric circulation, not only shifting regional precipitation (Xie et al. 2010, Sobel & Camargo 2011, Su et al. 2017) and the Hadley cells but also altering tropical cyclone activity (Vecchi & Soden 2007a, Knutson et al. 2008, Vecchi et al. 2008, Zhao & Held 2012).

The most recent comprehensive review of circulation change under warming is that by Schneider et al. (2010); it emphasizes theoretical analyses of water vapor latent heat release and the Hadley circulation, constrained by energetic, hydrological, and angular momentum balances, with idealized experiments between cold and warm extremes. The current review uses theories for atmospheric circulation to guide the diagnostic understandings of the changes that occur in observations and complex climate models, such as those in the Coupled Model Intercomparison Project (CMIP) phases 3 (Meehl et al. 2007) and 5 (Taylor et al. 2012).

The review covers existing and proposed theories; diagnosed mechanisms framed with spatial scales of the changes (e.g., global and regional, ocean and land); and future research directions. In Section 2, we summarize the theories for precipitation and the Walker and Hadley circulations and address their predictions. In Section 3, we first introduce the diagnostic slowdown of the tropical circulation and related physical mechanisms, then discuss regional precipitation change throughout the tropics. We review the unique mechanism that dominates precipitation change over oceans and influences the structure of the Hadley circulation in Section 4. In Section 5, we summarize the complicated change of land rainfall and introduce the direct radiative effect of $CO_2$. We discuss potential future research directions in Section 6 and follow with a summary in Section 7.

**Spatially uniform SST increase (SUSI):** an experiment with an atmosphere-only model forced by the same SST warming everywhere (usually 2 or 4 K)

**Coupled Model Intercomparison Project (CMIP):** a community-based standard experimental protocol for climate (change) model diagnosis, validation, intercomparison, documentation, and data access

**Intertropical convergence zone (ITCZ):** an area encircling Earth where the northeast and southeast trade winds converge within approximately ±20° of the equator

**Moist static energy (MSE):** combination of an air parcel's internal energy and energy for expansion, potential energy, and latent energy

**Geopotential:** magnitude of gravitational potential energy per unit mass, after removing the effects of rotation (e.g., centrifugal acceleration)

## 2. THEORETICAL FUNDAMENTALS AND PREDICTIONS

In the rising branches of the atmospheric circulation, air expands and cools, condensing water vapor, and the droplets grow, causing rainfall. The highly coupled climate system organizes the precipitation into zonally oriented bands. The tropical rain bands include the intertropical convergence zone (ITCZ), with precipitation that peaks near but to the north of the equator. Dry zones exist in the equatorial cold tongue and subtropical regions, with the exception of heavy monsoon rainfall over South and East Asia during summer.

Precipitation releases huge amounts of latent heat, which concentrates along these narrow rain bands in the tropics but also interacts strongly with the global atmospheric circulation. This can be explained with moist static energy (MSE), $h$, which is defined as $h = Lq + C_pT + \Phi$, where $q$ is specific humidity, $L$ is the latent heat of condensation per unit mass, $T$ is temperature, $C_p$ is the heat capacity of air at constant pressure, and $\Phi$ is the geopotential. Circulation is directly linked to gross moist stability, which can be interpreted as the MSE export from a tropospheric column by the mean circulation per unit of mean upward mass flux (Back & Bretherton 2006).

Regions with larger gross moist stability require smaller vertical circulation strength to export a given amount of MSE than do regions of smaller gross moist stability and therefore can be thought of as being more stable in a large-scale sense. Conversely, lower gross moist stability translates to

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larger upward velocities, and hence less stability, for a given export of MSE. Strong surface/low-level wind and moisture convergence over warm waters provides sufficient instability and latent heating to drive intense convection, so that the tropical precipitation bands are associated with SST of at least 26.5–27°C. The thermodynamic coupling between SST and deep convection also plays an important role in shaping rainfall change in global warming.

**Walker circulation:** a thermally driven equatorial zonal and vertical circulation, e.g., rising above the western Pacific, near the maritime Asian continent, and sinking over the eastern Pacific

## 2.1. Theories for Tropical Tropospheric Overturning Circulations

The tropical atmosphere is dominated by thermally driven overturning cells and monsoon circulations. As the monsoons have been systematically reviewed in the literature (An et al. 2015, Véspoli de Carvalho & Jones 2016), this section only discusses mechanisms driving and shaping the Walker and Hadley circulations.

**2.1.1. Walker circulation and Bjerknes feedback.** The equatorial Pacific Ocean features strong zonal asymmetry, characterized by the warm pool in the west and cold tongue in the east. The equatorial Atlantic Ocean has a similar thermal distribution, though the variations are weaker. The westward SST gradient confines deep convection mainly to the west warm pool, favoring an eastward gradient in sea surface pressure. As the Coriolis force is weak near the equator, the pressure gradient force drives prevailing easterly winds, generating oceanic equatorial upwelling via horizontal divergence induced by poleward Ekman currents. The westward wind forcing also tilts the sea surface to generate a balancing eastward pressure gradient force in the ocean, which shoals the eastern thermocline and allows cold water to be upwelled into the mixed layer and maintain the cold tongue. Such easterly winds at the surface, convection in the west, subsidence in the east, and westerly counterflow in the upper atmosphere compose the Walker circulation. This circular argument and interdependence between the ocean and atmosphere constitute the Bjerknes feedback (Bjerknes 1969) and are essential for explaining the slowdown of the Walker circulation in response to both El Niño and global warming (see the sidebar titled Bjerknes and Wind-Evaporation-SST Feedback).

**2.1.2. Hadley circulation and northward-displaced intertropical convergence zone.** On average, the tropics receive more solar radiation than the extratropics, and this energy imbalance is equilibrated by the heat transport via oceanic and atmospheric motion. The Hadley circulation is an important agent for poleward dry energy transport by the tropical atmosphere. This zonal-mean cell is located in both hemispheres and features ascent near the equator and subsidence in the

### BJERKNES AND WIND-EVAPORATION-SST FEEDBACK

The Bjerknes feedback is an ocean-atmosphere positive feedback shaping equatorial climate variability and change. For example, in the equatorial Pacific, relaxation of the easterly trade winds causes a flattening of sea level and thermocline, and hence a reduced upwelling-induced cooling of the eastern Pacific. This local warming effect decreases the westward SST gradients and hence further weakens the winds.

The wind-evaporation-SST feedback is an ocean-atmosphere interaction process shifting the intertropical convergence zone northward from the equator. A disturbance of SST to be warmer north of the equator causes cross-equatorial southerly winds. The Coriolis force then deflects the winds, decelerating the trade winds north of the equator and accelerating winds to the south. The evaporation of the northern tropics is weakened, thereby warming the local SST, and vice versa to the south. This positive feedback amplifies the initial disturbance.

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subtropics. The poleward flow in the upper troposphere is turned by the Coriolis force, forming strong westerly subtropical jets at the poleward edges of the Hadley cells, with the help of eddy momentum fluxes. In contrast, the air moves equatorward at the surface, and the Coriolis force results in easterly trade winds.

However, the rising branch of the Hadley circulation, known as the ITCZ, is not on the equator but to the north, despite the strongest solar insolation occurring at the equator. A wind-evaporation-SST feedback (Xie & Philander 1994) was suggested to explain this displacement (see the sidebar titled Bjerknes and Wind-Evaporation-SST Feedback). Assuming an SST perturbation, with positive anomalies to the north and negative to the south of the equator, the induced atmospheric pressure anomalies would drive a southerly cross-equatorial wind. The Coriolis would force wind anomalies to be easterly to the south and westerly to the north of the equator, increasing and decreasing the local easterly trade wind speed, and hence intensifying and reducing the corresponding evaporative cooling, respectively. This warms the SST to the north and vice versa, amplifying the initial perturbation. Thus, a positive wind-evaporation-SST feedback is formed to break the equatorial symmetry set by the annual-mean solar radiation, yet it does not favor either hemisphere. The following processes have been demonstrated to initiate the climatic asymmetry: asymmetric land-sea distribution, westward-propagating asymmetric Rossby waves, and SST-stratus feedback over the broad subtropical subsidence regions.

**2.1.3. Latitudinal boundary of the Hadley cell.** The Hadley circulation is thought to be caused by thermal gradients extending from the equator to the poles; however, the cells cease at around 30°N/S. Held & Hou (1980) explained this lateral restriction with simplified dry thermodynamics, by considering an axisymmetric flow and assuming the atmosphere is semifree of friction. First, the angular momentum conservation is used at the tropopause with the assumption of negligible vertical motion and friction, equivalent to a simplified zonal momentum equation. From this simplified momentum balance, the upper-level zonal wind can be determined as a function of latitude. The upper-level winds also define the average vertical wind shear, because the surface zonal wind is negligible due to surface friction. Then the meridional geostrophic balance, or the equivalent thermal wind balance, can be used to derive the meridional profile of tropospheric-mean potential temperature as a function of latitude and tropopause height.

The latitudinal integration of the tropospheric-mean potential temperature minus its radiative equilibrium value should be zero from the equator to the poleward boundary of the Hadley cell because the energy of the circulating air is conserved. At the poleward boundary, the actual and equilibrium potential temperatures should also be equal because of the below-discussed nature of radiative transfer. When an air parcel moves from the equator to the pole, it is cooler than it would be in radiative equilibrium. It is therefore warmed at first, and after it exceeds the equilibrium temperature, it starts to lose heat. After it reaches the equilibrium temperature again, there is no such energy source to heat it again, so the motion has to cease. This determines the latitudinal boundary of the Hadley circulation, which is proportional to the square root of the tropopause height and related to the shape of the equilibrium temperature.

Taking typical values observed from the current climate as input, the Hadley cell is constrained to terminate at 20–30°N/S and cannot extend all the way to the poles. The Hadley circulation flattens the meridional temperature profile by transporting heat poleward: High potential temperature is transported poleward in the upper troposphere, and cold air returns equatorward near the surface. This cools temperature at the equator and raises temperature at the poleward terminus of the circulation, compared to the radiative-convective equilibrium. Midlatitude eddies with traveling lows and highs also transport both sensible and latent heat poleward, helping to terminate the Hadley circulation in the subtropics.

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## 2.2. Future Changes Predicted by the Theories

The meridional Hadley circulation acts with the Walker circulation to regulate atmospheric temperature, humidity, cloudiness, and rainfall in the tropics and subtropics, with important implications for the hydrological cycle over ocean and land. In order to predict the changes in both circulations, it is necessary to extend the above theories to a warming climate. This section discusses a poleward expansion of the subsidence branch of the Hadley circulation, an energy constraint for slowdown of the tropical circulation, and a hypothetical pattern of future rainfall change.

**Greenhouse gas (GHG):** a gas in the atmosphere absorbing and emitting infrared radiation and warming Earth; examples include $H_2O$, $CO_2$, $CH_4$, $N_2O$, $O_3$

**Thermal wind relation:** vertical wind shear (variation of geostrophic wind with pressure) in proportion to horizontal temperature gradient with a direction paralleling the isotherms

**2.2.1. Poleward expansion of the tropics.** Although the inviscid theory of Held & Hou (1980) omitted two important dynamical factors, upper-level friction and moisture effect of latent heating, their simplified framework still predicts a poleward expansion of the Hadley cell. Increases in greenhouse gas (GHG) concentrations warm the troposphere and cool the stratosphere, resulting in a rise of the tropopause. This allows the circulating air of the Hadley cell to travel further, as tropopause height was shown to be important for its poleward limit. The connection between tropopause height and Hadley cell can be explained in the context of the thermal wind relation. An upward extension of the troposphere reduces vertical wind shear since the momentum conservation is hardly influenced. As a consequence of the thermal wind balance, the meridional gradient of potential temperature is reduced, extending poleward the latitude at which the potential temperature matches the equilibrium temperature. This is consistent with the slowdown of the tropical circulation associated with a flattened temperature gradient suggested by MASC (Ma et al. 2012).

More realistic theories that include the effects of midlatitude eddies give similar predictions with three dynamical mechanisms to explain global warming–induced expansion of the tropics more than Held & Hou (1980) would suggest. Lu et al. (2007) suggested that an increase in static stability and an associated reduction in baroclinicity in the subtropics (baroclinicity describes the misalignment between pressure and density gradients) cause eddy activity to retreat to higher latitudes and the Hadley cell to expand poleward. Changes in vertical wind shear could also contribute to the expansion.

The second mechanism involves upper-tropospheric baroclinic waves. With the rising tropopause in the subtropics, the phase speed of waves increases, weakening the waves' equatorward penetration and resulting in a poleward shift of the eddy momentum flux convergence and the position of the eddy-driven jet (Chen & Held 2007, Lorenz & DeWeaver 2007, Chen et al. 2008). For the third mechanism, Kidston et al. (2015) explained the tropical expansion from the standpoint of the stratosphere. They suggested that the stratospheric circumpolar westerly jet forming in winter has a coupled influence on tropospheric dynamics. Global warming induces a strengthening of the circumpolar jet, causing a poleward shift in the storm tracks and tropospheric jet stream.

Model diagnostics also suggest that radiative changes associated with clouds and water vapor affect the extent of the tropical boundaries. For example, Voigt & Shaw (2015) found that changes in tropical ice clouds contribute to an expansion of the tropics, while increased water vapor reduces the expansion, with significant intermodel uncertainty in the magnitude of these effects. Using a global model simulation, Wang et al. (2015) found a reduced meridional stream function and zonal winds over the tropics, as well as a poleward shift of the jet stream. They attributed the weakened and expanded tropical circulation to global redistribution of aerosol emissions from traditional industrialized countries to fast-developing Asia, which has caused a weakening of the meridional temperature gradient.

There is accumulating observational evidence confirming the theoretical prediction that the subsidence boundaries of the tropics in both hemispheres are expanding poleward. In a critical

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assessment, Lucas et al. (2014) identified five methodologies used to define the tropical expansion: tropopause-, satellite-, stream function–, jet stream–, and surface-based methods.

In summary, observations, model diagnostics, and theoretical predictions agree that the tropics should have expanded poleward, though there is uncertainty in the rate of the expansion, and there are multiple ways to understand it. The poleward extension of the subtropical dry zone with the poleward expansion of the tropics may bring drought conditions to certain regions that currently enjoy temperate weather (Seidel et al. 2008). Fu et al. (2006) demonstrated that the poleward shift leads to midlatitude warming and contributes to an increased frequency of drought in both hemispheres, including in many heavily populated regions. Therefore, reducing uncertainties in estimates and providing robust information for the expansion and its interpretation are important goals of future research.

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**2.2.2. Moist static energy constraint for tropical circulation.** A series of studies (Chou & Neelin 2004; Chou et al. 2009, 2013b; Chou & Chen 2010) used conservation of the aforementioned MSE to examine projected changes in the tropical circulation. MSE is imported into ascent regions through low-level moisture convergence and exported at upper levels through the horizontal divergence of high-MSE air. As there is a net flux of MSE into ascent regions from surface and radiative fluxes within those regions, this must be balanced by MSE export at upper levels. A number of mechanisms for circulation change were proposed under this framework, two of which (the dynamical rich-get-richer and upped-ante hypotheses) are described in Section 2.2.3.

A third mechanism predicts that the tropical circulation should weaken in global warming, with the effect of increasing depth of convection on the efficiency of net MSE export from tropical ascent regions (i.e., the gross moist stability). Under global warming, low-level temperature, moisture, and MSE import increase, which, therefore, has to be balanced by an increase of the MSE being exported at upper levels (assuming no large compensating changes in surface or radiative fluxes in ascent regions). This is achieved both by the enhanced warming of air at upper levels (discussed in Section 3.1.3) and by an increase in the height of convective outflow (as MSE increases with height in the upper troposphere).

In order for the tropical circulation to weaken, this increase in MSE at the level of convective outflow must not only balance the increase in low-level MSE but also overcompensate, so that the efficiency of net MSE export is increased and the same amount of MSE can be exported by a weaker circulation (increased gross moist stability). Chou et al. (2013b) showed that warming at upper levels alone is not enough to produce a sufficient increase in MSE export for the circulation to weaken, and that the height of convective outflow (depth of convection) must also increase for this to occur. Therefore, the weakening tropical circulation appears to be closely linked to the increased depth of convection, which is itself strongly related to changes in radiative emission by water vapor at upper levels (Hartmann & Larson 2002, Ingram 2010, Zelinka & Hartmann 2011).

**2.2.3. The rich-get-richer theory for regional rainfall change.** The wet-get-wetter concept was raised to describe how the patterns of evaporation ($E$) and precipitation ($P$) would change under global warming, including two related but slightly different hypotheses. The most commonly used version (Held & Soden 2006) proposed that in the (hypothetical) absence of circulation change, increased atmospheric water vapor implies an increased moisture transport from dry to wet regions, and hence an increased gradient of $P–E$. This leads to greater $P–E$ in wet regions and greater $E–P$ in dry regions, which appears to have been confirmed on very large averaged spatial scales by both observational (Allan et al. 2010, Durack et al. 2012) and modeling (Allan 2011) studies and

## WET-GET-WETTER AND WARMER-GET-WETTER

In the (hypothetical) absence of circulation change, increased atmospheric water vapor implies an increased moisture transport from dry to wet regions, and hence an increased precipitation gradient; i.e., rainfall increases in the core of existing rainy regions and decreases in current dry areas and at convective margins.

A general warming of the tropical troposphere raises the SST threshold for tropical convection, so that convective precipitation increases where SST warming exceeds the tropical mean and decreases where relatively weak warming exists. This SST pattern effect is achieved by adjusting the atmospheric circulation and corresponding surface divergence field, including two outstanding modes: an equatorial peak anchoring a local precipitation increase, and a meridional dipole with increased rainfall and weakened trade winds over the warmer hemisphere.

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is often described as the thermodynamic component of water cycle change (see the sidebar titled Wet-Get-Wetter and Warmer-Get-Wetter).

The alternative version is a dynamical feedback on this increased atmospheric moisture (Neelin et al. 2003, Chou & Neelin 2004), described as the rich-get-richer or anomalous gross moist stability mechanism. Originally based on experiments with a coupled ocean-atmosphere-land model of intermediate complexity (Neelin & Zeng 2000, Zeng et al. 2000), this mechanism was also examined in the CMIP3 projections (Chou et al. 2009). Under global warming, the moistened boundary layer (i.e., the lowest part of the atmosphere, which is sensitive to the presence of Earth's surface) reduces gross moist stability and consequently enhances convection and precipitation in convective regions. If it applied throughout the tropics, a consequence of this dynamical feedback would be enhanced convergence into ascent regions, and hence a strengthened tropical circulation. As overall the circulation actually weakens (see Section 3.1), this dynamical rich-get-richer mechanism cannot be the dominant controller of tropical circulation change in response to warming, though it could be important locally.

To explain the shift of the maximum rainfall increase from the core rainy regions (see Section 3.2.1), Chou & Neelin (2004) raised the additional upped-ante hypothesis, arguing that a warmer troposphere increases the boundary-layer moisture threshold (the ante) for convection. This could result in reduced rainfall in the margins of convective regions, if the increase in moisture advection into these regions is insufficient to meet the raised convective threshold. However, rainfall does not reduce in all convective margins, meaning that the upped-ante applies selectively in some regions but not others (Chadwick et al. 2013). Indeed, this convective threshold-raising theory is part of the warmer-get-wetter view (Ma & Xie 2013; discussed in Section 4.1.1), and is applicable only on the margins where the SST increase is lower than the tropical mean. Otherwise the ante needed to start convection remains roughly the same.

**Boundary layer:** the lowest part of the atmosphere, directly influenced by the surface, with turbulent wind, temperature, and moisture and strong vertical mixing

## 3. LARGE-SCALE CIRCULATION AND REGIONAL PRECIPITATION CHANGES ACROSS THE TROPICS

Here we extend Section 2.2.2 by introducing diagnostic evidence and mechanisms raised to explain the large-scale slowdown of the tropical circulation. We then compare the spatial pattern of the weakening with the precipitation gradient increase theoretically hypothesized in Section 2.2.3. As the universal dynamical and thermodynamic components of regional rainfall change across the tropics, they are found to significantly offset each other.

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### 3.1. Weakening of the Large-Scale Atmospheric Circulation

A weakening of the tropical circulation and an associated increase in the average atmospheric residence time of water vapor are robust projections of all climate models. This section discusses observational evidence for the circulation slowdown and physical mechanisms to explain it.

**3.1.1. Observed and simulated weakening of tropical circulation.** Since the mid-nineteenth century there has been a weakening trend in the Indo-Pacific zonal sea level pressure gradient, which was suggested by Clarke & Lebedev (1996) as evidence of a weakening Walker circulation. Vecchi et al. (2006) demonstrated that model simulations could accurately reproduce the observed trend in sea level pressure gradient, but only when they included anthropogenic forcing, indicating that the weakening Walker circulation was primarily due to human activities. Recent work has raised more uncertainty about the weakening of the surface winds, with observed global and tropical increases in wind speed during 1987–2006 (Wentz et al. 2007, Ma et al. 2016) and a positive trend in the global evaporation since the late 1970s (Yu 2007). A new data set, which uses wave height to compensate for errors due to a changing level of wind observations (Tokinaga & Xie 2011), suggests that there has indeed been a Walker circulation slowdown over the past 60 years (Tokinaga et al. 2012), though overall twentieth-century trends of surface temperature and pressure in the Pacific remain uncertain across different data sets (Solomon & Newman 2012).

This observational evidence supports model projections of further weakening of the tropical atmospheric circulation during the twenty-first century (Tanaka et al. 2004). Held & Soden (2006) found a reduction in the amount of convection in CMIP3 models, which is also present in the more recent CMIP5 projections (Chadwick et al. 2013, Su et al. 2014). Vecchi & Soden (2007b) further reported a reduction in the frequency of strong updrafts and an increase in the frequency of weak updrafts.

Several complementary hypotheses have been proposed to explain the slowing of the tropical tropospheric circulation, and these are outlined here. This large-scale circulation response to general SST warming can be viewed from a variety of different perspectives, such as the differing rates of change in water vapor, lapse rate, precipitation, and radiative cooling (see the sidebar titled Thermodynamic, Radiative, and Dynamical Hypotheses for Tropical Circulation Weakening). In fact, these diagnostic results are likely to all be part of the same overall mechanism, with unsolved questions remaining besides the convective depth–related constraint (Section 2.2.2).

#### THERMODYNAMIC, RADIATIVE, AND DYNAMICAL HYPOTHESES FOR TROPICAL CIRCULATION WEAKENING

The global-mean water vapor and vertical gradient of air temperature increase at ~7% $K^{-1}$ of surface warming, but the fractional changes in global-mean precipitation and net longwave radiative cooling are only 1–2% $K^{-1}$ based on model projections. This contrast indicates a weakening of the circulation at a rate of ~5% $K^{-1}$ in both convective and subsidence regions.

Mean advection of stratification change (MASC) is a dynamical mechanism proposed to explain the weakening of the tropospheric circulation throughout the tropics in climate change. Following moist adiabats, tropics-wide warming increases with height and stabilizes the atmosphere. This leads to relative cooling of the air column in ascending regions due to anomalous cold advection of low-level air, and relative warming in subsidence regions due to warm advection. This reduces air temperature and pressure gradients between ascent and descent regions, opposes the climatological circulation, and slows it down as an adiabatic forcing.

**Clausius-Clapeyron equation:** a relation giving the slope of the pressure-temperature coexistence curve of a phase transition in terms of entropy and volume changes across the transition

**Moist adiabatic lapse rate:** the rate at which atmospheric temperature decreases with an increase in altitude, including latent heating by condensation of water vapor

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**3.1.2. Thermodynamic relation on convective mass-flux change.** Held & Soden (2006) combined the Clausius-Clapeyron equation for the saturation vapor pressure with an approximate equation for precipitation,

$$P = Mq, \qquad 1.$$

to diagnose how the tropical circulation should change under warming. Here $P$ is precipitation, $M$ is the vertical convective mass flux, and $q$ is a typical boundary-layer specific humidity. This can be understood simply as "what goes up must come down," and it shows that any increase in boundary-layer moisture ($q$) must result in increased precipitation unless it is balanced by a decrease in convective mass flux ($M$).

For CMIP3 A1B simulations, the global-mean water vapor increase was diagnosed to be $\sim$7% $K^{-1}$, in close agreement with the expected increase in saturation vapor pressure from Clausius-Clapeyron. The fractional change in global-mean precipitation, however, is only 1–2% $K^{-1}$ based on projections from coupled climate models. This contrast indicates that the circulation in the convective regions has to slow down at a rate of $\sim$5% $K^{-1}$ of the global ocean surface warming. This reduction in time-mean convective mass flux appears to manifest itself in a decrease in the frequency and/or duration of convective events (Sun et al. 2007), which is robustly seen in climate model and high-resolution cloud resolving model simulations (Singh & O'Gorman 2013), rather than a reduction in the updraft intensity of individual storms.

Yet the relation itself is only part of the explanation of the slowdown, with the dynamical mechanism that constrains the interaction between precipitation and radiation to have a muted response still lacking.

**3.1.3. Radiative relation on subsidence change.** A time-mean reduction of upward motion over the entire rising branch of the Walker circulation was reported by Knutson & Manabe (1995), accompanied by a similar weakening of the easterly trade winds throughout the tropical Pacific. This result seemed surprising at the time, given increased condensation and enhanced precipitation over the western Pacific warm pool region, which is now known to result from the ability of the atmosphere to hold more moisture. They attributed the weakening of the circulation to an imbalance in the subsidence regions important for regulating the overall distribution of the tropical greenhouse effect (Pierrehumbert 1995, Williams et al. 2009): Net radiative cooling of the troposphere in these regions does not increase as quickly as the vertical gradient of air temperature, as discussed in more detail below.

Temperature follows an approximately moist adiabatic lapse rate almost uniformly throughout the tropics, a quasi-uniformity known as the weak temperature gradient approximation (e.g., Bretherton & Sobel 2002). In ascent regions convection causes the atmospheric temperature profile to be moist adiabatic, and this vertical profile is then propagated to the rest of the tropics by fast equatorial waves. A global SST increase shifts the temperature profile throughout the whole atmospheric column, with air warming increasing with height according to the curvature feature of the moist adiabats. Recent work (O'Gorman & Singh 2013) suggests that the change in the atmospheric temperature profile in response to surface warming may be more accurately represented by a transformation involving an upward shift. Nevertheless, both descriptions result in enhanced warming at upper levels.

In the energy budget for subsidence regions, radiative cooling is mainly balanced by adiabatic warming via large-scale descent. Therefore, any change in radiative cooling must be balanced by some combination of change in lapse rate and change in circulation strength, which together determine the magnitude of the change in adiabatic warming. As models predict that the lapse rate

decreases more than the relatively weak enhancement in radiative cooling, the vertical motion in descent regions has to weaken to maintain the balance.

The thermodynamic and radiative relations use similar formulations between the convective and subsidence regions, respectively. This can be explained by an energy balance argument that latent heating of the troposphere (i.e., precipitation) increases as fast as net tropospheric longwave radiative cooling (1–2% $K^{-1}$), as the two are diagnosed to approximately balance (Allen & Ingram 2002). However, the amount of latent heat release demanded by the surface energy budget (Pierrehumbert 2010) can be achieved through many different atmospheric adjustments rather than radiative cooling (Pierrehumbert 1999). As a result, the latent heating and radiative cooling may be only weakly coupled, unless there exists a more subtle collective behavior involving the interaction of large-scale dynamics with radiation and convection. The solar energy absorbed at the surface (Le Hir et al. 2009) can be used as a constraint for evaporation to limit precipitation, but at the current climate status this constraint is insufficient to achieve the muted responses, since at least a 5% $K^{-1}$ increase can be yielded in a column radiative-convective model (Pierrehumbert 2002). Thus, the mechanisms that constrain the increase in radiative cooling have still not been convincingly explained.

**3.1.4. A dynamical mechanism associated with lapse-rate change.** Also related to lapse-rate change is the MASC mechanism of Ma et al. (2012). As discussed in Section 3.1.3, global SST warming leads to a tropics-wide warming that increases with height in the atmosphere. This stabilization results in a further effect on total column temperature through vertical advection of the change in lapse rate by the mean circulation. In contrast to the thermodynamic and radiative relations, which apply respectively to convective and subsidence regions, MASC is applicable throughout the tropics.

Upper-level air warms more than lower-level air, so this leads to relative cooling of the air column in ascending regions due to upward advection of cooler low-level air and relative warming of the column in subsidence regions due to downward advection of warmer upper-level air. This MASC advective effect distorts isotherms of air warming, and the effect is similar with or without spatial variation of the SST increase. In this way, vertical advection tends to reduce air temperature and pressure gradients between ascent and descent regions. This can be viewed as an adiabatic forcing opposing the climatological circulation, with the effect of slowing it down.

The MASC effect has an analytical expression and can be diagnosed from general circulation models and applied to a linear baroclinic model to examine the effect of such geographically uneven heating on atmospheric circulation (Watanabe & Kimoto 2000). The individual effect of MASC significantly weakens tropical thermal-driven circulations, including the Walker and Hadley cells and monsoon winds (Ma & Yu 2014b, Qu & Huang 2016), and reduces the meridional air temperature gradient and hence the tropical wind shear through thermal wind relation. As a purely dry effect, MASC is able to offset the enhanced latent heating due to moisture increase, because vertical temperature advection nearly balances diabatic heating in the long-term-mean large-scale tropical atmospheric dynamics. In addition, global warming features a pronounced tropical-mean SST warming four times larger than the spatial patterns, resulting in significant atmospheric stabilization and hence the MASC effect. In order for the linear baroclinic model to properly reproduce circulation change in a fully coupled model under GHG forcing, this effect has to be explicitly assigned, since it is not included by default. In contrast, for internal variability, such as El Niño, spatial variations in SST anomalies are 40% greater than the mean warming, so that stratification is less affected and latent heating can be used to drive the linear model effectively without MASC.

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**Representative Concentration Pathways (RCPs):** four GHG concentration (not emissions) trajectories adopted by the Intergovernmental Panel on Climate Change for its *Fifth Assessment Report* in 2014

Furthermore, since the climatological circulation is relatively similar between models and scenarios, as are the tropical-mean moist adiabats, the magnitude of MASC is also fairly independent of model physics and forcing scenarios. **Figure 1** provides evidence for this argument in the form of an empirical orthogonal function analysis of the MASC forcing among 76 simulations under three scenarios, including CMIP3 A1B and CMIP5 Representative Concentration Pathways (RCPs) 6.0 and 8.5, corresponding to medium and high GHG emission scenarios. Dominating 68.3% of the total variance, the first leading mode (**Figure 1*a***) shows significant spatial dependence on climatological pressure velocity, which indicates that the intermodel/interscenario variability is primarily in the magnitude of the circulation change as a whole, rather than in local differences due to model details and radiative forcing. A high correlation of 0.92 between the first principal component against tropical-mean (40°S–40°N) SST increase suggests that the strength of the MASC effect is mainly dependent on the magnitude of spatially uniform warming in each model (**Figure 1*b***).

## 3.2. Tropics-Wide Changes in Regional Precipitation

In addition to the global-mean warming effects, regional climate change is very important from a practical standpoint, and robust information is urgently needed (Xie et al. 2015). This section describes the observed regional circulation and rainfall changes in response to increased GHG concentrations, then discusses the offsetting mechanisms of change in the whole tropics. In addition to GHG forcing, the effects of which we examine here, changes in aerosol concentrations have been linked to large regional changes in circulation and the hydrological cycle. These aerosol-circulation interactions are not discussed, as they are reviewed by Lee et al. (2014).

**3.2.1. What do the observations tell?** Lau & Wu (2007) analyzed global precipitation products and found a positive trend along the equatorial oceans (5°S–5°N), but negative trends over the Indo-Pacific warm pool and central Africa. Zhou et al. (2011) found similar patterns, with the maximum rainfall increase occurring at 5°N, and interpreted the result as a strengthening of the ITCZ, consistent with the wet-get-wetter view. Allan et al. (2010) also examined trends during 1988–2008 (earlier data were considered less reliable), averaged separately over all ascending and all descending areas of the tropics. They found an increasing trend in wet ascent regions and a decreasing trend in dry descent regions, also consistent with a large-scale wet-get-wetter, dry-get-drier thermodynamic view of rainfall change. However, the maximum rainfall increase is actually shifted equatorward from the ITCZ's core, suggesting the importance of dynamical processes. In situ sea surface salinity observations (Curry et al. 2003, Durack & Wijffels 2010) also show trends consistent with an apparent acceleration of the hydrological cycle (enhanced rainfall in the tropics and enhanced evaporation in the subtropics), with patterns similar to large-scale CMIP5 predictions (Durack et al. 2012), though the magnitude is higher in the observations.

At very large averaged scales over land, there is evidence of wet seasons getting wetter and dry seasons drier (Chou et al. 2013a). However, at impacts-relevant regional scales over land, observed rainfall trends do not support a wet-get-wetter, dry-get-drier paradigm (Greve et al. 2014, Roderick et al. 2014). Hegerl et al. (2015) summarized these observed hydrological cycle changes in the context of the existing theories and predictions, e.g., wet-get-wetter, muted precipitation response, and intensified weather extremes. However, they suggested that although observations show robust evidence for theoretically derived and numerically predicted changes, uncertainties from the small signal-to-noise ratio of natural variability and limitations of short, discontinuous, and inhomogeneous observational data sets pose serious difficulties for determining the anthropogenic contributions.



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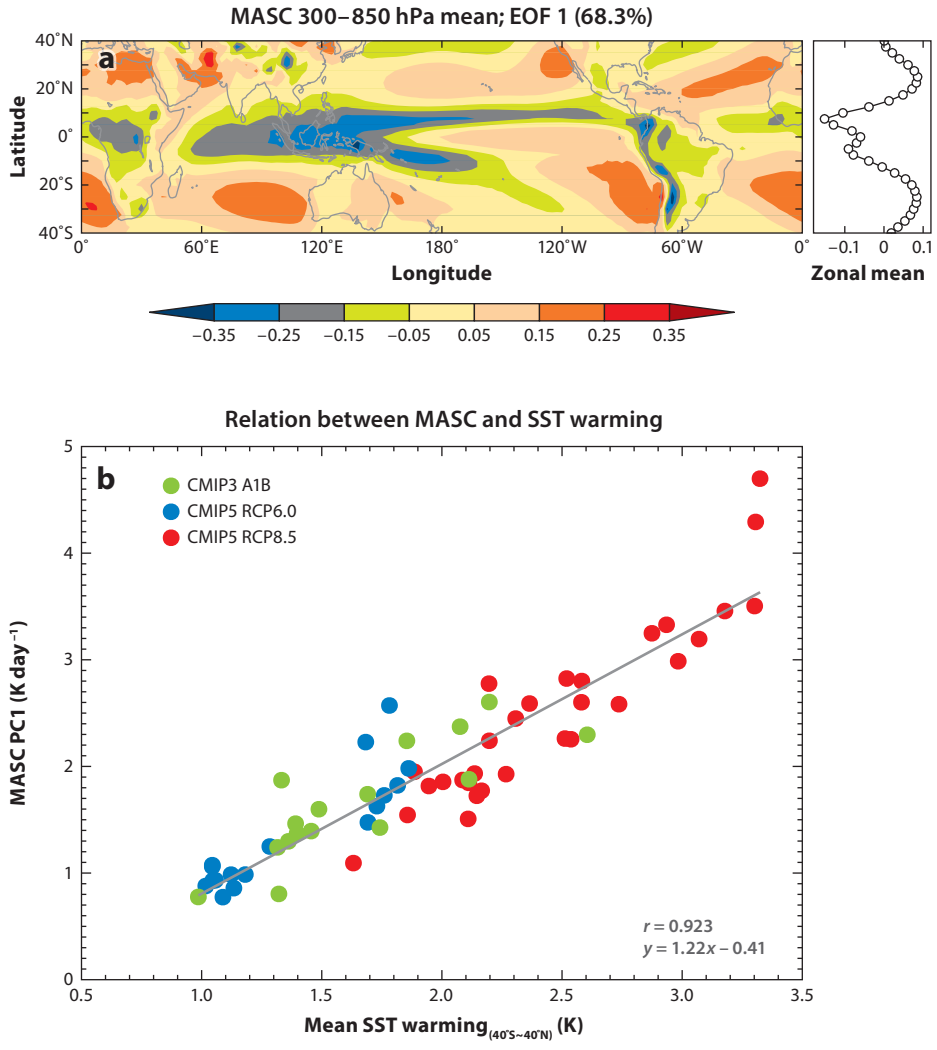



**Figure 1**

First leading mode of the intermodel empirical orthogonal function (EOF) analysis for the vertical-averaged mean advection of stratification change (MASC) forcing calculated with 76 simulations for three scenarios: the Special Report on Emissions Scenarios A1B (22) of Coupled Model Intercomparison Project (CMIP) phase 3 and the Representative Concentration Pathways (RCPs) 6.0 (23) and 8.5 (30) of CMIP5. Spatial mode 1 (panel *a*) shows its pattern (nondimensional) dependence only on climatological vertical velocity, and a scatter between tropical-mean sea surface temperature (SST) warming and the first principal component (PC) (panel *b*) indicates its magnitude dependence only on the general warming extent.

**3.2.2. Decomposition of precipitation change mechanisms.** A number of studies have used moisture and energy budgets to examine the regional response of circulation and precipitation change to GHG forcing and to decompose the total response into components associated with various mechanisms. Often the total precipitation change is partitioned into changes associated with thermodynamic (atmospheric moisture increases in response to warming) and dynamical

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(circulation) processes. Dynamical and thermodynamic changes interact to produce the total pattern of rainfall change, where the thermodynamic component usually represents a rich-get-richer term due to moisture increases.

The first such decomposition (Chou & Neelin 2004) used the MSE and moisture budgets to analyze regional rainfall change and proposed a number of mechanisms that could drive regional rainfall change. These are the dynamical rich-get-richer hypothesis (see Section 2.2.3), which could lead to increased rainfall in ascent regions; the increased depth of convection (Section 2.2.2), which could weaken the circulation and decrease rainfall; and the upped-ante mechanism (Section 2.2.3), which could cause rainfall decreases on certain margins of ascent regions.

More recent decompositions have used moisture (Seager et al. 2010, Bony et al. 2013, Chadwick et al. 2013) or dry static energy (Muller & O'Gorman 2011, Richardson et al. 2016) budgets to understand rainfall and circulation change. These studies found that at regional scales, the pattern of rainfall change is determined more by dynamical circulation changes than by thermodynamic moisture increases, e.g., subtropical rainfall reduction due to poleward shifts of the storm tracks (Scheff & Frierson 2012). In fact, the spatial correlation between the patterns of present-day rainfall and future change is low across future CMIP5 projections (Chadwick et al. 2013), which would not be the case if these changes were dominated by a rich-get-richer response to moisture increases. This can be explained by dynamical precipitation decreases, associated with the weakening circulation, tending to oppose the rich-get-richer pattern (Seager et al. 2010).

**3.2.3. Wet-get-wetter versus the regional weakening of circulation.** On the tropical-mean spatial scale, the circulation is diagnosed to weaken, and this must also be reflected in local circulation changes. Of the mechanisms described in Section 3.1, only MASC (Ma et al. 2012) provides a prediction of how this overall weakening manifests itself on a regional scale. Chadwick et al. (2013) used this to explicitly separate dynamical rainfall changes into a component corresponding to the weakening circulation and a residual associated with spatial shifts in convection. Under this formulation, the pattern of thermodynamic rainfall increases is strongly anticorrelated with rainfall decreases caused by the weakening circulation, resulting in a strong cancellation between the two terms.

Because MASC involves the mean vertical motion acting on the changes in spatially averaged stratification, its pattern is negatively proportional to the climatological circulation. **Figure 1*a*** shows anomalous cooling in the convective regions over the Indo-Pacific warm pool and ITCZ, and warming in the subtropical subsidence centers. **Figure 2** presents the effect of this forcing pattern in a linear baroclinic model, producing a fractionally uniform decrease in upper-level divergence over the warm pool. This represents a weakening of the Walker circulation, which in the mean is characterized by strong upper-level divergence (and lower-level convergence) over the warm pool. The uniformity in the decrease is stronger at upper rather than lower levels, possibly due to the physical effects of friction and orography or artifacts in the velocity potential calculation due to interference from orography.

Therefore, the prediction of MASC is that regions of strongest present-day vertical motion will experience the greatest future weakening, i.e., a wet-get-drier effect on regional precipitation. As part of the dynamical effect, this would oppose and mitigate the wet-get-wetter mechanism, increasing the relative importance of other dynamical processes. Indeed, the overall pattern of rainfall change is dominated by the pattern of shifts in convective regions (Chadwick et al. 2013), which could be driven by any mechanism that affects the locations where convection preferentially occurs. The balance of these processes differs between ocean and land: Certain dynamical mechanisms are more important over the former, while others dominate over the latter, so we now consider such details separately in the following two sections.



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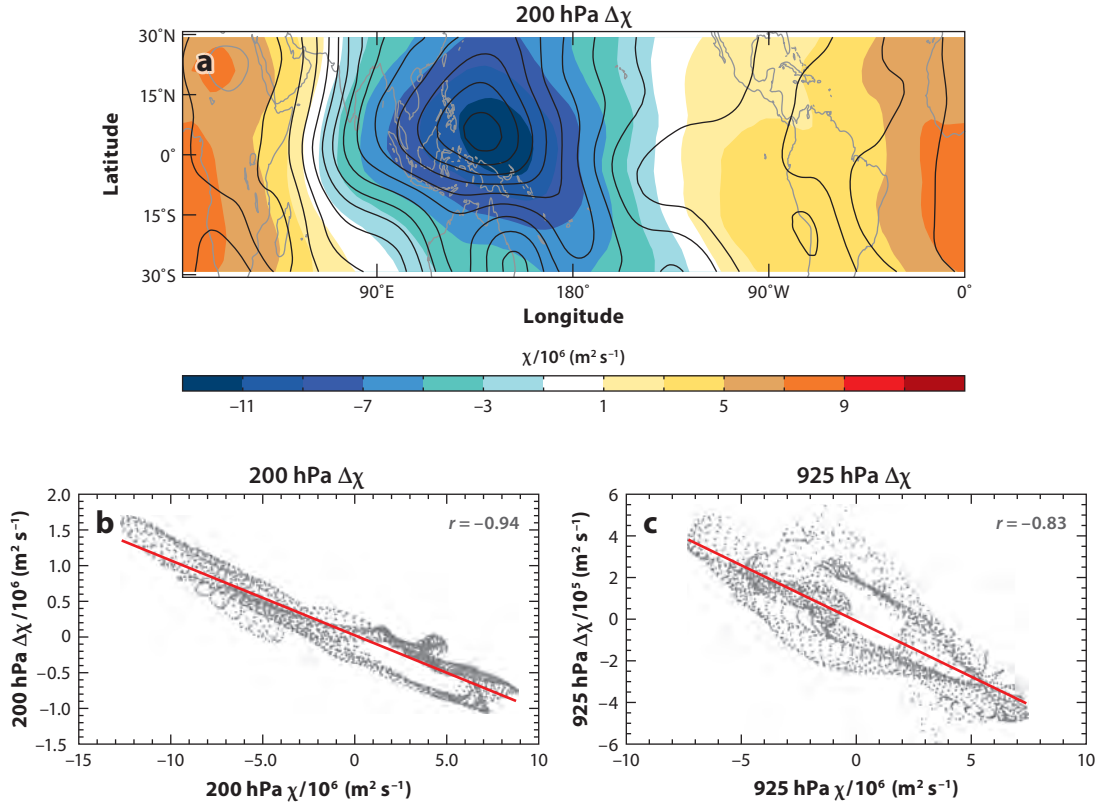



**Figure 2**

Local weakening of circulation induced by mean advection of stratification change forcing (**Figure 1*a***) in a linear baroclinic model. Represented with velocity potential ($\chi$), panel *a* shows the horizontal distribution of change (*contours*) over climatology (*color*) at the 200-hPa level, and scatterplots indicate their negative linear relationship at (*b*) 200 hPa and (*c*) 925 hPa.

## 4. CHANGES RELATED TO SEA SURFACE TEMPERATURE PATTERNS

As mentioned in Section 2, SST is a predominant driver shaping atmospheric circulation and precipitation. Here we show that dynamical mechanisms associated with SST patterns are important for changes in oceanic rainfall and the Hadley circulation.

### 4.1. Circulation and Precipitation Changes over the Oceans

An equatorial peak of precipitation change was frequently illustrated in previous studies. Because the global zonal-mean view makes it difficult to distinguish the rainfall patterns from the purely thermodynamic prediction, it was considered wet-get-wetter. However, both the full two-dimensional pattern and the zonal mean of the projected rainfall change are correlated weakly with climatological rainfall in the tropics (Ma & Xie 2013), showing the importance of mechanisms other than the wet-get-wetter.

**4.1.1. The warmer-get-wetter paradigm.** The SST and tropospheric temperature anomalies seem to lack spatial variations in comparison with the mean warming across the tropics (e.g.,



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Neelin et al. 2003); however, the warmer-get-wetter mechanism suggests that such SST pattern change is influential in regional rainfall and circulation change, with increased convection and ascent over regions where SSTs warm the most (see the sidebar titled Wet-Get-Wetter and Warmer-Get-Wetter). It was first proposed from a gross moist instability estimation by Xie et al. (2010), in which the instability was defined as the difference in MSE between the ocean surface and upper troposphere. In the tropics, the upper-tropospheric temperature increase varies spatially by <0.3 K, with its gradients flattened by fast equatorial wave adjustments (Sobel et al. 2001, Bretherton & Sobel 2002). Consequently, a general warming of the tropical troposphere raises the SST threshold for tropical convection (Johnson & Xie 2010), and moist instability and thus convective precipitation increase where the SST warming exceeds the tropical mean and decrease where relatively weak warming exists. The upped-ante theory (Chou & Neelin 2004) is the moisture alternative of this threshold-raising effect in the convective margins where the SST increase is weak.

Xie et al. (2010) decomposed SST warming into a tropical mean and spatial deviations, and Ma & Xie (2013) examined the warmer-get-wetter effect induced by the SST patterns with large CMIP3 and CMIP5 ensembles. Two SST patterns stand out: an equatorial peak (Liu et al. 2005) anchoring a local precipitation increase, and a meridional dipole mode with increased rainfall and weakened trade winds over the warmer hemisphere. These SST patterns were found to be important for explaining both the ensemble mean distribution and intermodel variability of rainfall change over the tropical oceans (Ma & Xie 2013, Ma & Yu 2014a). As commonly seen in the tropics (e.g., Back & Bretherton 2006), this effect is shown by a moisture budget analysis to involve strong positive feedback between atmospheric circulation and convection.

**4.1.2. Combined mechanism for tropical oceanic rainfall change.** The warmer-get-wetter mechanism is not the only mechanism controlling the regional oceanic rainfall response, but it is a very strong mechanism for spatially shifting convection over the oceans (Section 3.2). Chadwick et al. (2013) and Ma & Xie (2013) suggested that the regional oceanic rainfall change is given by the following approximate expression, derived from Equation 1 ($P$, $M$, and $q$ are precipitation, vertical convective mass flux, and boundary-layer specific humidity, respectively):

$$\begin{aligned} \delta P/P &= \delta M_{\text{shift}}/M + (\delta M_{\text{weak}}/M + \delta q/q) \\ &= \alpha\, T^* + \beta\, \overline{T} \end{aligned}, \qquad 2.$$

where $\bar{T}$ is tropical-mean SST warming, $T^*$ denotes SST patterns, $\alpha$ and $\beta$ are parameters describing the corresponding strength of the response, and the subscripts "weak" and "shift" refer to the circulation slowdown and spatial shift of precipitation, respectively. In percentage form, this approximates the full mechanism controlling the regional precipitation response over the tropical oceans.

The global SST warming contains a uniform increase and spatial patterns. If $T^*$ is zero, Equation 2 represents the rainfall response in the SUSI. As mentioned in Section 2.2.3, the general warming moistens the atmosphere to cause wet-get-wetter ($\delta q/q$), and at the same time reduces the tropical circulation ($\delta M_{\text{weak}}/M$) to mitigate the moisture effect. This results in a muted precipitation response ($\beta \bar{T}$). However, even if $T^*$ is moderate in comparison to $\bar{T}$, it can dominate circulation and precipitation change over the SUSI effect in the coupled climate system, since the factor $\alpha = 44\%\ \text{K}^{-1}$ is much larger than $\beta = 2\%\ \text{K}^{-1}$, according to a linear fit of spatial distribution between $T^*$ and $\delta P/P$ in the CMIP3 ensemble mean (Ma & Xie 2013). The SST patterns shift the atmospheric circulation ($\delta M_{\text{shift}}/M$), causing a warmer-get-wetter ($\alpha\, T^*$) rainfall response.

As moisture budget analyses (Seager et al. 2010, Chadwick et al. 2013) show, wet-get-wetter is the thermodynamic component of regional precipitation change, and circulation slowdown and the



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SST pattern effects are the dynamical components. Because the circulation weakening contributes a wet-get-drier effect, counteracting the wet-get-wetter effect, the atmospheric circulation change associated with SST patterns (warmer-get-wetter) dominates the total rainfall redistribution over the tropical oceans.

## 4.2. Structural Changes of the Hadley Circulation

Besides a poleward expansion, more complicated changes occur to the circulation structure within the Hadley cells. They include a plausible shift of the ITCZ and an inhomogeneous intensity change reflecting a competition between circulation weakening and the SST pattern effect.

**4.2.1. Interhemispheric energy balance and the intertropical convergence zone shift.** Constituting the upward branch of the Hadley cell, the ITCZ is the strongest convection belt in the tropics. A growing body of research has examined the link between the interhemispheric energy balance and the position of the zonal-mean ITCZ, including its response to forcing. These studies have often used idealized models to examine the underlying processes more clearly, and the extent to which the same effects determine the response of more complex fully coupled models to forcing is an open question.

Kang et al. (2009) used a comprehensive atmosphere model coupled to a slab mixed layer ocean (i.e., ocean dynamics were suppressed) to study the effects of the extratropics on the position of the ITCZ. They imposed a cross-equatorial heat flux beneath the ocean's surface mixed layer to cool the northern extratropics and warm the southern extratropics, and this induced a southward shift in the ITCZ. This displacement can be understood as a compensation by the atmospheric energy transport in response to the imposed oceanic heat flux in the tropics: In response to the southward shift of heat in the ocean, the ITCZ shifts southward and the northern Hadley cell strengthens, resulting in an increase in the northward atmospheric heat transport. The magnitude of the ITCZ shift was found to be sensitive to cloud feedback.

Using fundamental energy constraints to analyze the responses of nine CMIP3 slab ocean model simulations to a doubling of $CO_2$, Frierson & Hwang (2012) found that differences in extratropical clouds control the diversity of ITCZ responses. Positive feedbacks involving water vapor and high clouds in the tropics were shown to reinforce the initial ITCZ responses. Seo et al. (2014a) found that high-latitude forcing causes a larger shift in the ITCZ than forcing in the tropics. Equivalent simulations without cloud and water vapor feedbacks, however, showed a weaker ITCZ shift when the forcing was farther from the equator, emphasizing the importance of radiative feedbacks in their experiments. Related to this, Fučkar et al. (2013) suggested that the ocean's meridional overturning circulation causes the ITCZ to be located in the Northern Hemisphere, where deep water is produced, since the Southern Hemisphere circumpolar flow forces northward oceanic heat transport.

**4.2.2. Inhomogeneous change of the Hadley cell intensity.** Section 3.1 discussed the slowdown of the tropical tropospheric circulation; however, robust observational evidence of this slowdown is found only for the Walker cell. In contrast, strengthening trends in the Hadley cell over the past few decades are reported based on prevailing reanalysis data sets, though this could be internal multidecadal variability rather than a forced trend (Quan et al. 2004; Tanaka et al. 2004; Mitas & Clement 2005, 2006). Future climate model projections suggest stronger future weakening of the Walker than the Hadley circulation (Gastineau et al. 2009, Ma & Xie 2013). Current theories on large-scale circulation change are insufficient to explain these different responses, so that regional effects have to be accounted for (Xie et al. 2015). Equation 2 suggests that the SST

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pattern effect turns out to be important for the oceanic rainfall change, with the partial offset between an increase in humidity and a decrease in circulation. For circulation change, humidity is irrelevant, so that the circulation weakening is as important as the SST pattern effect. Together, they are the dominant dynamical components of circulation change (Section 4.1.2) over the ocean.

Ma & Xie (2013) found that while the weakening circulation acts to slow down both the Hadley and Walker circulations, as shown in SUSI experiments, their different responses in the coupled models could be explained by the influence of SST pattern change. Weakening of the Walker cell is well established because the influence of the SST pattern effect is either weak (because of zonal symmetry in CMIP3) or acting to enhance the slowdown (due to the eastern Pacific peak in SST warming in CMIP5). However, the impact of the weakening circulation on the Hadley cell is more influenced by the SST patterns. They are additive in some geographical regions and opposing in others, resulting in robust weakening north of the equator but weak and highly uncertain changes near and south of the equator. This is due to the latitudinal dependence of the SST patterns, representing a combination of the equatorial peak and meridional asymmetry. However, the seasonal changes in the Hadley circulation (Seo et al. 2014b) are dominated by changes in the latitudinal potential temperature gradient across the subtropics, including the robust (slight) weakening of the Hadley cell during boreal winter (summer).

Outside of the tropics, the pattern of future SST change appears to have overall little impact on the response of the atmospheric circulation and, in turn, on the resulting changes in precipitation. This is due to the insensitivity of Rossby wave generation to the changes in near-equatorial upper-level divergence (He et al. 2014).

## 5. DIRECT $CO_2$ EFFECT AND LAND RAINFALL CHANGE

Extra absorption of upwelling longwave radiation by increased GHG concentrations enables the troposphere to be moderately warmed even without SST increases. This warming (with a maximum at about 700 hPa) can reduce the radiative cooling of the troposphere and thus the convective mass flux, slowing the tropical circulation and weakening global- and tropical-mean precipitation (Allen & Ingram 2002, Sugi & Yoshimura 2003, Yang et al. 2003, Lambert & Webb 2008, Dong et al. 2009, Andrews et al. 2010, Bala et al. 2010, Cao et al. 2012, O'Gorman et al. 2012, Kamae et al. 2015). This is known as the direct radiative effect of $CO_2$ on precipitation change, and it could also strengthen and shift poleward the midlatitude westerly winds (Deser & Phillips 2009). In addition, the land surface is free to warm in response to the increased downwelling longwave radiation from the extra $CO_2$, and this can destabilize the atmosphere (Giannini et al. 2013); increase flow from ocean to land; and enhance convection, rainfall, and evaporation over land but suppress precipitation over ocean. This section first evaluates the former, then discusses the latter along with other mechanisms of land rainfall change.

### 5.1. The Direct $CO_2$ Radiative Effect

The direct $CO_2$ effect is commonly diagnosed from atmosphere model experiments where $CO_2$ concentrations are increased but SSTs held constant. This causes problems of interpretation, because keeping SST fixed provides an infinite energy source/sink, voiding energetic consistency.

Bony et al. (2013) found that by the end of the twenty-first century, under a high GHG forcing scenario, approximately one-quarter to one-third of the projected mean tropical circulation change is independent of global-mean surface warming and can thus be attributed to the direct $CO_2$ effect. Because aqua-planet experiments (all land removed and replaced with ocean) show tropical-mean results that are consistent with simulations that include land, they rejected a major contribution



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from land warming toward these regional circulation changes. Chadwick et al. (2014) followed up this work and confirmed that the tropical-mean circulation change is substantially affected by the direct radiative effect and is therefore to some extent independent of global-mean temperature change. However, regional patterns of rainfall change are dominated by surface warming patterns, including both SST pattern change and land-sea temperature contrast change. They suggested that future regional rainfall changes should be studied primarily with coupled models.

He et al. (2014) and He & Soden (2015) examined these mechanisms in more detail and confirmed that mean SST warming, the direct $CO_2$ effect (including land warming), and SST pattern change all play roles in regional circulation change, though the effects of SST pattern change are mainly limited to the tropics. Deser & Phillips (2009) investigated the relative importance of direct atmospheric radiative and observed SST forcing on observed global atmospheric circulation trends during December through February of 1950–2000. They suggested that both drive distinct responses that contribute about equally to the full circulation pattern trend and are approximately additive and partially offsetting. Direct radiative effects drive the strengthening and poleward shift of the midlatitude westerly winds in the Southern Hemisphere (and to a lesser extent over the Atlantic-Eurasian sector in the Northern Hemisphere), while SST trends (especially in the tropics) act to intensify the Aleutian low and weaken the Walker circulation.

Model simulations show differences between global precipitation changes during the twentieth and twenty-first centuries (Thorpe & Andrews 2014), which can be attributed to the changing balance between the influence of SST warming and direct atmospheric absorption. The historical precipitation changes little despite increasing SSTs, because large direct effects from $CO_2$ and black carbon oppose the surface warming–induced precipitation increase, while in future scenarios the importance of these direct effects declines and the SST increase dominates in both ensemble mean and uncertainty and in both global mean and spatial patterns.

## 5.2. Circulation and Precipitation Changes over Land

Circulation changes over tropical land are generally less well understood than those over the oceans, and they have often been analyzed on a region-by-region basis. There is a large body of research describing circulation changes in monsoon regions (An et al. 2015, Véspoli de Carvalho & Jones 2016), and we do not attempt to describe monsoon or other region-specific mechanisms here in as much detail. We instead discuss a number of mechanisms proposed to drive circulation and rainfall changes across the tropical land.

**5.2.1. Amplified land warming and effect on the circulation.** Surface temperatures warm more over land than over the oceans in response to GHG forcing, and this is associated with pressure, circulation, and rainfall changes (Bayr & Dommenget 2013). This enhanced land warming is true even at equilibrium (Joshi et al. 2008), so while the smaller heat capacity and faster warming response of the land surface compared to the oceans play a role in the land amplification in transient warming scenarios, the heat capacity is not the main cause of the effect.

The land surface is warmed by increased net downwelling radiation due to the direct response of the atmosphere to $CO_2$ forcing, but also by the response of the atmosphere and land surface fluxes to remote SST warming (Giannini 2010), and in fact, the remote response is larger (Chadwick 2016). Compo & Sardeshmukh (2009) attributed the recent worldwide land warming to the SST warming rather than the direct GHG effect. Atmospheric model simulations of the past half-century successfully reproduced most of the land warming when prescribed with observed SST, but without GHG changes. Hydrodynamic-radiative teleconnections were suggested to be the primary mechanism for warming the land, mainly through moistening and warming the air over

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## HYDRODYNAMIC-RADIATIVE TELECONNECTIONS AND THE BOWEN RATIO

Hydrodynamic-radiative teleconnections are a thermodynamic process in climate change enabling the oceans warmed by a greenhouse gas increase to influence land warming. Horizontal advection, diffusivity, and/or wave actions from the ocean can increase moisture and temperature of the air over land, both of which enhance the downward longwave radiation at the surface. This may account for most of the land warming.

The Bowen ratio is the ratio of surface-to-atmosphere sensible heat divided by latent heat. The surface sensible heat is generated by the difference between surface temperature and near-surface air temperature; the surface latent heat is generated by the difference of saturation mixing ratio at surface temperature and the actual specific humidity of near-surface air. The Bowen ratio is an indicator of the type of surface (<1 over surfaces with abundant water supplies), e.g., tropical oceans (<0.1), rainforests (0.1–0.3), temperate forests and grasslands (0.4–0.8), semiarid landscapes (2.0–6.0), and deserts (>10.0).

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land by the ocean and increasing the downward longwave radiation at the land surface (see the sidebar titled Hydrodynamic-Radiative Teleconnections and the Bowen Ratio).

As described in Section 3.1.3, SST warming leads to a quasi-moist-diabatic response of the vertical temperature profile (diabatic, the opposite of adiabatic, refers to a thermodynamic change with exchange of heat into or out of the system). In the free troposphere this warming is homogenized across the tropics by equatorial waves (e.g., Bretherton & Sobel 2002), leading to an almost uniform horizontal temperature distribution at mid- and upper levels. Over land, the vertical temperature profile in the boundary layer and at the surface must adjust so as to maintain a smooth vertical temperature profile (Joshi et al. 2008). As moisture supply is limited over land but not over the oceans, the present-day balance between latent and sensible surface heat fluxes (the Bowen ratio; see the sidebar titled Hydrodynamic-Radiative Teleconnections and the Bowen Ratio) varies between the two, with a larger proportion of sensible heating over land. Equivalently, land has less water to evaporate than the oceans, so that more energy goes into warming the surface (Sutton et al. 2007). So in general the mean lower-tropospheric lapse rate over land is larger (more dry-adiabatic and less moist-adiabatic) than over the oceans. For a given temperature in the midtroposphere, the surface temperature over land would therefore be higher than the equivalent surface temperature over the oceans. In a similar way, the response of land surface temperatures to SST-driven free-tropospheric warming is larger than the SST warming itself, as sensible heating increases more relative to latent heating over land than it does over the oceans (Joshi et al. 2008, Lambert et al. 2011).

**Free troposphere:** atmosphere above the boundary layer, where winds are approximately geostrophic (parallel to isobars), and usually nonturbulent or only intermittently turbulent

**Equivalent potential temperature:** the temperature an air parcel would reach if brought adiabatically to sea level, with all its water vapor condensed

**Relative humidity (RH):** the ratio of the partial pressure of water vapor to the equilibrium vapor pressure at a given temperature

This argument can be formulated in an alternative way as a requirement that the change in equivalent potential temperature is uniform over both land and ocean (Byrne & O'Gorman 2013). As atmospheric moisture increases less over land than over the oceans, this leads to the requirement that land warming must be greater than ocean warming. In this framework the enhanced land warming is due to a combination of the lower relative humidity (RH) over land than over ocean in the present-day warming, and further decreases in future land RH.

Clearly this land temperature adjustment is coupled to evaporation and humidity change, and it is also closely related to circulation change. Enhanced warming over land extends from the surface through the lower troposphere, and this leads to pressure anomalies and circulation changes (Bayr & Dommenget 2013). Additionally, SST-driven general warming in the free troposphere has the effect of stabilizing the atmosphere over land and suppressing convection (Joshi et al. 2008, Giannini et al. 2013), which reduces rainfall and couples with evaporation changes. These two

effects may be in competition with each other, as the surface warming should drive a sea breeze–type anomalous circulation, bringing air from the oceans to land, whereas suppression of convection and convective heating over land would tend to reduce the flow from ocean to land. It is possible that different circulation changes are seen at different vertical levels. Overall, the net effect of tropical-mean SST warming is to reduce convection and rainfall over land (Bony et al. 2013, Giannini et al. 2013, Chadwick et al. 2014, He et al. 2014).

**5.2.2. The direct $CO_2$ effect–associated land warming.** Land heating in response to increased downwelling longwave radiation also drives a circulation response. This can be relatively simply interpreted as being due to destabilization of the atmosphere over land (Giannini et al. 2013), with increased flow from ocean to land and increased convection, rainfall, and evaporation over land (Dong et al. 2009, Cao et al. 2012, Biasutti 2013, Bony et al. 2013). In this case, pressure changes from land warming and increased convective heating would combine to drive increased sea-land flow. Ackerley & Dommenget (2016) examined the effects of increasing land surface temperatures while keeping SSTs and $CO_2$ concentrations constant and found that the pattern of circulation and rainfall change is very similar to that of experiments where $CO_2$ is increased but SSTs fixed. This suggests that the main influence of direct $CO_2$ heating on regional circulation change is via land heating, rather than atmospheric stabilization.

The competing effects of direct local $CO_2$ warming and remote uniform SST warming on circulation and rainfall change (Giannini et al. 2013) can be examined in idealized atmosphere-only model experiments, where one aspect is fixed and the other increased (Bony et al. 2013, Chadwick et al. 2014, He & Soden 2015, Richardson et al. 2016). As expected, the two effects generally oppose each other, and in many regions the total response is a relatively small residual of the two larger terms.

**5.2.3. Relative humidity decreases over land.** Over the oceans, surface RH is predicted to remain approximately constant under warming (Held & Soden 2000, Schneider et al. 2010), with only small RH increases projected to inhibit latent heat increase (Richter & Xie 2008). Over land, however, substantial future decreases in RH are projected in many regions (O'Gorman & Muller 2010, Byrne & O'Gorman 2015). As on long timescales almost all moisture over land originates from the oceans, it has been suggested that this RH drying over land is due to amplified land warming, as advection of moisture from the relatively less-warmed oceans is unable to keep up with the increased moisture-holding capacity of the warmer air over land (Rowell & Jones 2006, Simmons et al. 2010, O'Gorman & Muller 2010).

RH decreases can be driven by circulation, precipitation, and evaporation changes, but they are also themselves a driver of circulation and water cycle change. This is most obvious in the relationship between RH and boundary-layer moisture, which controls the thermodynamic change in future precipitation through local moisture availability (Chou et al. 2009, Seager et al. 2010, Chadwick et al. 2013) and potentially also via advection of humidity gradients by the circulation (Byrne & O'Gorman 2015).

There is also a possible dynamical influence of RH decreases on circulation change, due to the influence of RH on cloud-base height (Fasullo 2012). Lower RH could lead to higher cloud-base heights, potentially suppressing convection, and intermodel correlations are seen in many regions between the magnitude of RH changes and shifts in the regions of convection (Chadwick 2016), though this does not necessarily imply causality.

The coupling between humidity, water cycle, and circulation change makes the chain of causality difficult to entangle, and targeted idealized modeling experiments may be the best way to gain a better understanding of these mechanisms. Overall, the circulation and water cycle response over

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tropical land is likely to consist of a combination of responses to the various mechanisms listed here, with different balances of change emerging in different regions. If these mechanisms can be better understood, it should be possible to reduce the currently substantial intermodel uncertainty in future simulations by evaluating the most important physical processes for any given region between models and observations. Other mechanisms may also be important in some regions, but their description is underdeveloped; they are discussed in Section 6.3.

## 6. DIRECTIONS FOR FUTURE RESEARCH

This section identifies promising fields for future research, in order to constrain uncertainties and provide more reliable climate projections. We suggest that fundamental understandings are needed on the mechanisms shaping the changes in SST, the ITCZ, atmospheric circulation, tropical cyclone activity, and land rainfall.

### 6.1. Combining Sea Surface Temperature Pattern Change and Intertropical Convergence Zone Shifts

It is noteworthy that the extratropical forcing of the ITCZ shift associated with oceanic energy transport has been studied mostly with idealized models (see Section 4.2.1). However, it is also important to investigate the projections by comprehensive coupled models for which ITCZ shifts have to be consistent with SST pattern changes and tropical air-sea interactions. As mentioned in Section 4.1.1, one major feature of the SST warming is the interhemispheric asymmetry, with stronger warming to the north contributing to weakening of the Hadley circulation there. Indeed, Friedman et al. (2013) showed an increasing trend of the south-to-north gradient in the observed SST (<0.8 K) since 1980 and inferred a northward shift of the ITCZ. They showed that this SST gradient trend is simulated by the CMIP5 models and is projected to continue increasing significantly in the twenty-first century.

**Figure 3** uses SST and rainfall as indicators to show the projected change in position of the tropical Pacific ITCZ in the twenty-first century. Simulated by 19 CMIP5 RCP4.5 models, the SST maximum not only shifts northward but also expands toward the equator, and the southward expansion is more significant than the northward expansion. This translates to an ITCZ widening at ~0.5° latitude per century toward both hemispheres, consistent with maximum SST warming on the equator and in the northern subtropics. The precipitation change supports this interpretation, showing equatorward expansion, consistent with more climatological precipitation toward the equator and the abovementioned energy theories.

As previous studies suggested, in addition to the impact of the atmosphere and ocean energy transport on the ITCZ position, local dynamical and thermodynamic ocean-atmosphere interactions associated with the SST patterns may also be important. For instance, the northward ITCZ expansion may be attributed to a wind-evaporation-SST feedback (Ma & Xie 2013), and the equatorward expansion to the reduced damping rate with lower mean surface evaporative cooling of the colder SST (Xie et al. 2010, Lu & Zhao 2012), enhanced by the Bjerknes feedback. Moreover, linking SST patterns to the ITCZ change may provide an alternative view of the tropical climate change dynamics. At least in the Northern Hemisphere, the widening of the ITCZ can be a part of the tropical expansion. The MASC effect also acts to reduce the surface air temperature gradients and may influence the SST through evaporation adjustment. Combined, these mechanisms can put the atmospheric circulation change in a new perspective and link the uncertainties in climate projection to the biases in simulating the climatology.

The current consensus considers the ITCZ shift and SST patterns separately, though they are likely to be two sides of the same coin. Long et al. (2016) attempted to link the interhemispheric



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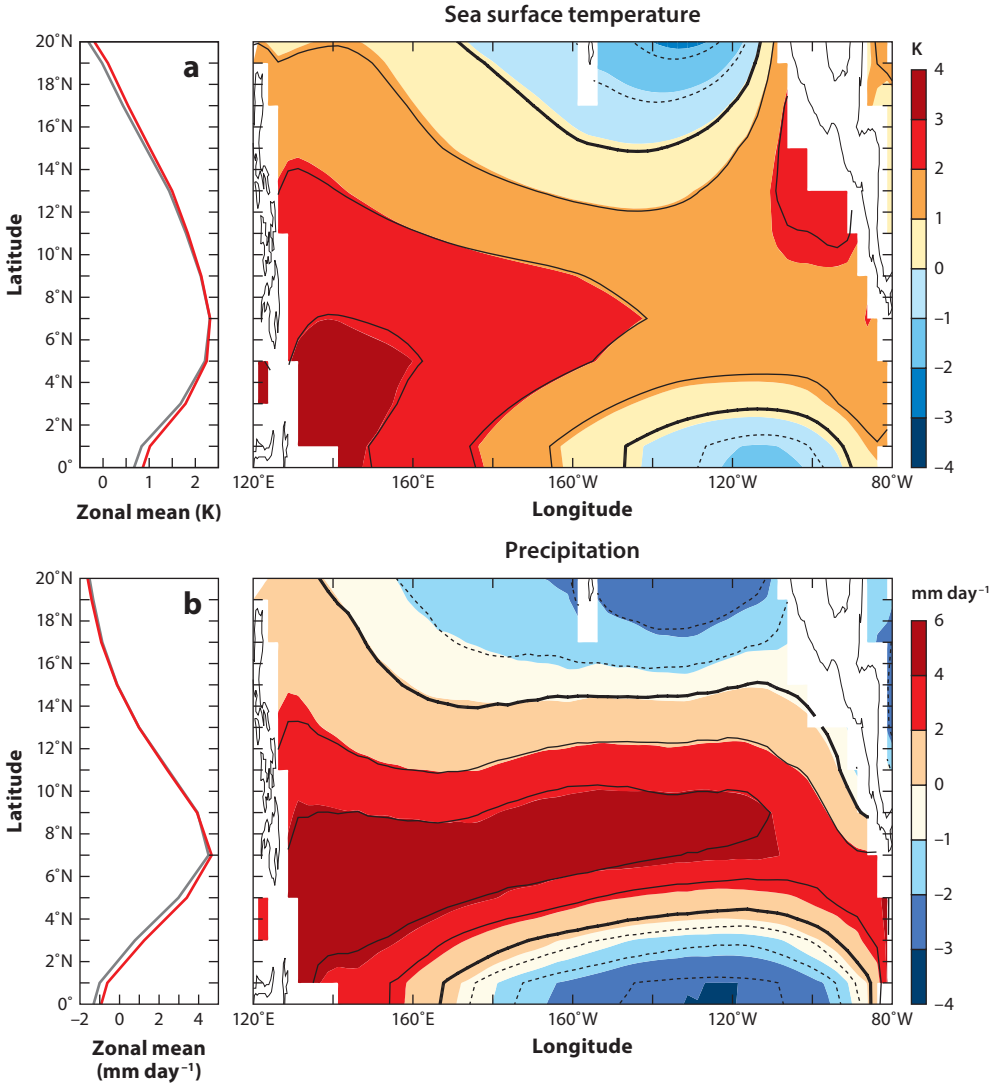



**Figure 3**

Ensemble and time mean of (*a*) sea surface temperature (K) and (*b*) precipitation (mm day$^{-1}$) illustrating change of the Pacific intertropical convergence zone in the twenty-first century as simulated by 19 Coupled Model Intercomparison Project phase 5 (CMIP5) models along Representative Concentration Pathway 4.5. The current climate is calculated as the average during 2006–2015 (*contour*; *black* for zonal mean) and the future climate as the average during 2089–2098 (*color*; *red* for zonal mean). The patterns show the deviations from the tropical Pacific (30°S–30°N, 120°E–80°W) mean.

energy balance that drives the ITCZ change to the SST patterns; however, more work is needed to fundamentally understand how the SST patterns emerge. We suggest a synthesized budget analysis combining both sides—surface fluxes and ocean transport. Besides short- and longwave radiation, surface latent and sensible heat fluxes are worth examining, since they involve the influence from winds, RH, and air-sea instability.

## 6.2. Impact of Circulation Change on Tropical Cyclones

Tropical cyclones constitute a significant part of seasonal rainfall in the tropics, and environmental conditions influence their genesis and development, primarily through vertical wind shear and SST (DeMaria 1996). As global warming weakens the tropical circulation, wind shear should lessen, increasing intense tropical cyclone activity and intensity (Knutson et al. 2008, Bender et al. 2010). Nevertheless, Vecchi & Soden (2007c) reported enhanced shear in the main hurricane development regions in the tropical Atlantic and East Pacific, supported by past observations (Wang & Lee 2008, Wang et al. 2008). This enhancement was shown to be caused by SST patterns (weaker warming in the north relative to the equator) and would not suggest a strong anthropogenic increase in hurricane activity. In addition, the relative SST change was reported to be important (Vecchi et al. 2008, Wang & Lee 2008) to provide static energy and moisture to penetrate the increased threshold for convection.

We have illustrated that the reduction of vertical wind shear associated with MASC is nearly model and scenario independent (**Figure 1**), and that much of the uncertainty in hurricane responses should come from the SST patterns (Zhao & Held 2012). We suggest an intermodel analysis comparing the influence of large-scale slowdown and SST pattern mechanisms on hurricane change for both the past and the future as an important direction for future study. Two key measures should be investigated: maximum potential intensity for development (Emanuel 1999), which is primarily controlled by SST patterns, and genesis potential index (Camargo et al. 2007) or the recently developed cyclone genesis index (Bruyère et al. 2012) for genesis, which considers both SSTs and wind shear.

## 6.3. Balance of Processes Driving Circulation Change over Land

As described in Section 5.2, there are several processes that could potentially lead to circulation changes over land, and the balance of these differs among regions. Understanding more about these processes is crucial for narrowing the large uncertainty in climate model projections of future regional circulation and water cycle change.

A new set of atmosphere-only time-slice experiments (the piSST experiments based on the preindustrial control coupled simulation and a4SST experiments on the run with abruptly quadrupling $CO_2$) will be included in the contribution of the Cloud Feedbacks Model Intercomparison Project phase 3 to CMIP6 (Webb et al. 2017), designed to analyze the processes that drive regional climate change. These numerical experiments will isolate the individual responses of climate models to direct $CO_2$ forcing, uniform SST warming, pattern SST warming, the plant physiological effect, and sea ice change and should provide a much greater understanding of the processes that drive change and uncertainty across coupled climate models for any given land region, as is done for the ocean (Section 4.1.2).

# 7. SUMMARY

This article reviews the past ~20 years of progress in understanding tropical atmospheric circulation change cause by global warming and its impacts on the hydrological cycle. We summarize the theories of atmospheric circulation and their predictions, present the diagnosed physical mechanisms underlying large-scale and regional changes in observations and climate models, and suggest a few promising future directions for improved climate projections. We decompose the global-mean warming effects from the tropics-wide regional climate change, considering both thermodynamic and dynamical components that drive circulation and precipitation changes. Various mechanisms for precipitation change are discussed separately over oceans and land, with related changes in the Hadley circulation and the direct $CO_2$ radiative effect, respectively.



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Theoretical predictions of the atmospheric circulation change, and its effects on the hydrological cycle, have been identified, including a poleward expansion of the Hadley cell, the slowdown of the tropical circulation, and a wet-get-wetter pattern for precipitation. Diagnostic results then suggest that the latter two significantly offset each other, giving rise to an SST pattern dominance of regional precipitation change over the oceans. As a result, robust weakening is found only for the Walker circulation and the Hadley circulation in the northern subtropics, with great uncertainty for the latter near and to the south of the equator. These changes are also related to shifts of storm tracks and the ITCZ.

Dynamical mechanisms such as the MASC effect and the warmer-get-wetter paradigm were proposed for these equilibrium changes. More complicated mechanisms are likely to be important over land, and this is an important area of ongoing research effort. Many other challenges also remain for future research, including the tropical cyclone environment and ITCZ shifts. In general, a more fundamental understanding of the pattern formation of SST, circulation, and rainfall will be crucial to narrowing uncertainty in future climate projections.

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## SUMMARY POINTS

1. Significant progress has been made in understanding the climate change–induced responses of the tropical atmospheric circulation and hydrological cycle.
2. These changes can be separated into large-scale responses and regional changes, showing connections between sea surface temperatures, winds, precipitation, and energy transport.
3. Theoretical predictions include a poleward expansion of the Hadley cell, the slowdown of the tropical circulation, and a wet-get-wetter trend for precipitation.
4. The large-scale weakening of the tropical circulation exerts a regional dynamical impact on precipitation change, by partially offsetting the thermodynamic wet-get-wetter effect.
5. This allows local dynamical processes to dominate regional circulation and precipitation changes, e.g., the warmer-get-wetter effect over ocean and the land surface warming and $CO_2$ effects over land.
6. Impacts of circulation weakening and sea surface temperature patterns are comparable for the Hadley cell change and competitive near and south of the equator, which may be related to a shift of the intertropical convergence zone.

## FUTURE ISSUES

1. A combined understanding of changes in the sea surface temperature patterns and the intertropical convergence zone may provide an alternative view for global warming pattern formation, so that uncertainties in climate projections can be traced back to the biases in current climate simulations.
2. Uncertainty in environmental change of tropical cyclones needs to be better understood.
3. A budget analysis for land rainfall change is needed to examine the balance of multiple processes for various land regions.

## DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

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**Errata**

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LETTERS
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# Distinct global warming rates tied to multiple ocean surface temperature changes

Shuai-Lei Yao[1,2], Jing-Jia Luo[3]*, Gang Huang[1,2,4]* and Pengfei Wang[1,5]

**The globally averaged surface temperature has shown distinct multi-decadal fluctuations since 1900[1–4], characterized by two weak slowdowns in the mid-twentieth century and early twenty-first century and two strong accelerations in the early and late twentieth century. While the recent global warming (GW) hiatus has been particularly ascribed to the eastern Pacific cooling[5,6], causes of the cooling in the mid-twentieth century and distinct intensity differences between the slowdowns and accelerations remain unclear[7,8]. Here, our model experiments with multiple ocean sea surface temperature (SST) forcing reveal that, although the Pacific SSTs play essential roles in the GW rates, SST changes in other basins also exert vital influences. The mid-twentieth-century cooling results from the SST cooling in the tropical Pacific and Atlantic, which is partly offset by the Southern Ocean warming. During the recent hiatus, the tropical Pacific-induced strong cooling is largely compensated by warming effects of other oceans. In contrast, during the acceleration periods, ubiquitous SST warming across all the oceans acts jointly to exaggerate the GW. Multi-model simulations with separated radiative forcing suggest diverse causes of the SST changes in multiple oceans during the GW acceleration and slowdown periods. Our results highlight the importance of multiple oceans on the multi-decadal GW rates.**

The Paris Climate Conference in December 2015 has agreed to pursue efforts to limit the global warming (GW) to no more than 1.5 °C above pre-industrial levels by 2100 (ref. 9). The observed annual-mean globally averaged surface temperature (GST) has already risen about 0.89 °C (0.69–1.08 °C, 90% confidence interval)[10] during the period 1900–2012 (Fig. 1 and Supplementary Table 1), leaving room for just another 0.5–0.6 °C warming relative to today for the rest of the twenty-first century. While the early twenty-first-century GW hiatus indicates that the relentlessly increasing greenhouse gases (GHGs) may not always boost the GST warming simultaneously[6], it is yet unclear whether the recent hiatus has surely ended and how strongly GW will resume in the next decades. In addition, it is found that the GW rate since 1900 has experienced a pronounced multi-decadal fluctuation, characterized by two accelerations in the early and late twentieth century and two slowdowns during the mid-twentieth century and the early twenty-first century[1,2] (Fig. 1). Correspondingly, the globally averaged land surface air temperature shows similar multi-decadal modulations. The distinctive multi-decadal fluctuation undoubtedly affects the estimation of centennial mean GW rate[11,12]. Improved understanding of these GW accelerations and slowdowns could help provide a better estimate of the future GW rate for the Paris Climate Agreement.

The estimated rates of the multi-decadal GW accelerations and slowdowns are subject to the time interval selection[13], which is often based on an arbitrary decision[14]. Here, the GW acceleration and slowdown periods are objectively determined according to three thresholds: a maximum magnitude of statistically significant trends in observed GST, relatively low uncertainty among different observations (represented by error bars in Supplementary Fig. 1), and a high percentage of global areas where the merged land air temperature and ocean surface temperature has a statistically significant trend (Supplementary Fig. 1 and Methods). Our approach identifies two periods with maximum accelerations of the GW during 1908–1945 and 1975–1998 and two periods with maximum slowdowns during 1940–1976 and 2001–2012 (Fig. 1).

It is found that observed GST trends during the two acceleration periods are statistically significant across four different observational data sets, whereas those during the two slowdown periods are less significant (Supplementary Table 1), especially for the early twenty-first-century slowdown. The latest slowdown does not show a consistently negative GW rate among the different observations and is statistically insignificant (Supplementary Table 1), probably owing to a relatively short period for the trend estimation. Despite this, the recent GW slowdown has received tremendous attention and has been examined extensively[5,6,8]. The magnitudes of the GW rates during the two slowdown periods are considerably smaller than those during the two acceleration periods. This difference has been explained as being due to the naturally occurring decadal/multi-decadal climate variability halting the steady GW rate attributed to the continuously increasing GHGs emissions[15]. For instance, the recent slowdown has been tied to the equatorial Pacific sea surface temperature (SST) cooling[5,6], which is underpinned by the large impacts of Interdecadal Pacific Oscillation (IPO) on the GST[16,17]. The IPO's impacts are similar but not identical to those exerted by El Niño[18].

Consistent with previous findings[5,6], SST warming (cooling) occurs in the tropical Pacific during the acceleration (slowdown) periods (Supplementary Fig. 2). The SST change differences between the slowdown and acceleration periods bear a close resemblance to the pattern of the cold phase of the IPO[19] or a La Niña-like decadal/multi-decadal variability[20]. This indicates the possible importance of internal climate variability in modulating GW[5,16]. However, notable differences in global SST changes exist among the acceleration and slowdown periods (Fig. 2). While the SST shows a similar warming in the tropical Pacific between the two acceleration periods (Fig. 2a,c), the equatorial central-eastern Pacific cooling during 2001–2012 is far stronger than that during 1940–1976 (Fig. 2b,d). Its magnitude is also greater than those

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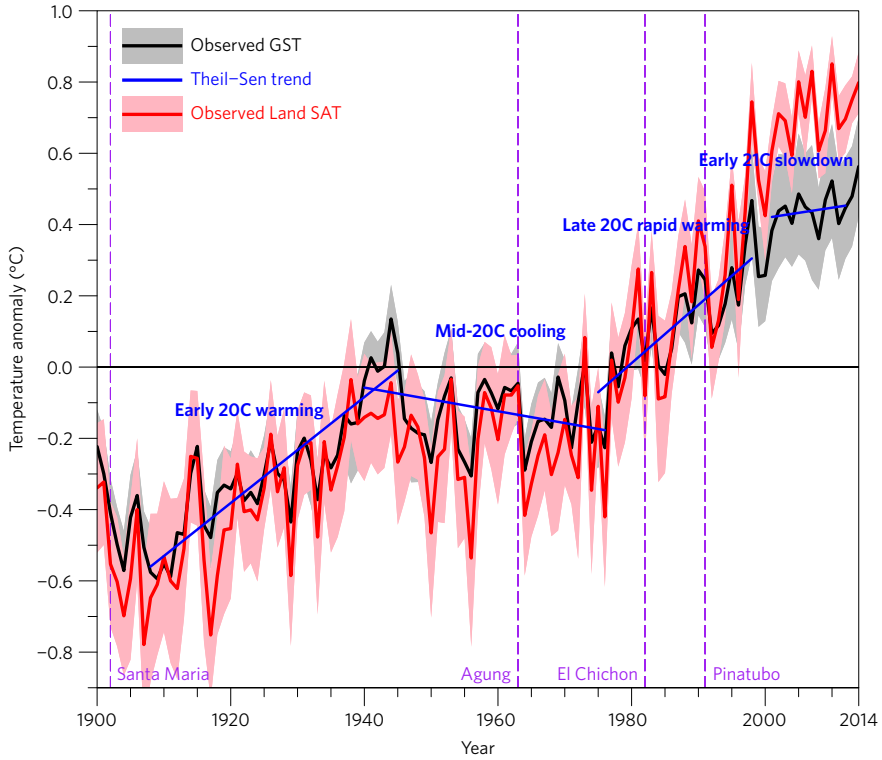



**Figure 1 | Observed annual-mean globally averaged surface temperature (GST) and land surface air temperature (SAT) anomalies.** Anomalies are relative to 1961–1990 climatology. The grey-shaded and pink-shaded areas represent one standard deviation of the differences among the observed GST and SAT, respectively. The solid blue lines indicate the Theil–Sen trend estimates for the early twentieth-century warming period (1908–1945), the mid-twentieth-century cooling period (1940–1976), the late twentieth-century rapid warming period (1975–1998) and the early twenty-first-century slowdown period (2001–2012), respectively. The vertical purple lines indicate the dates of major volcanic eruptions.

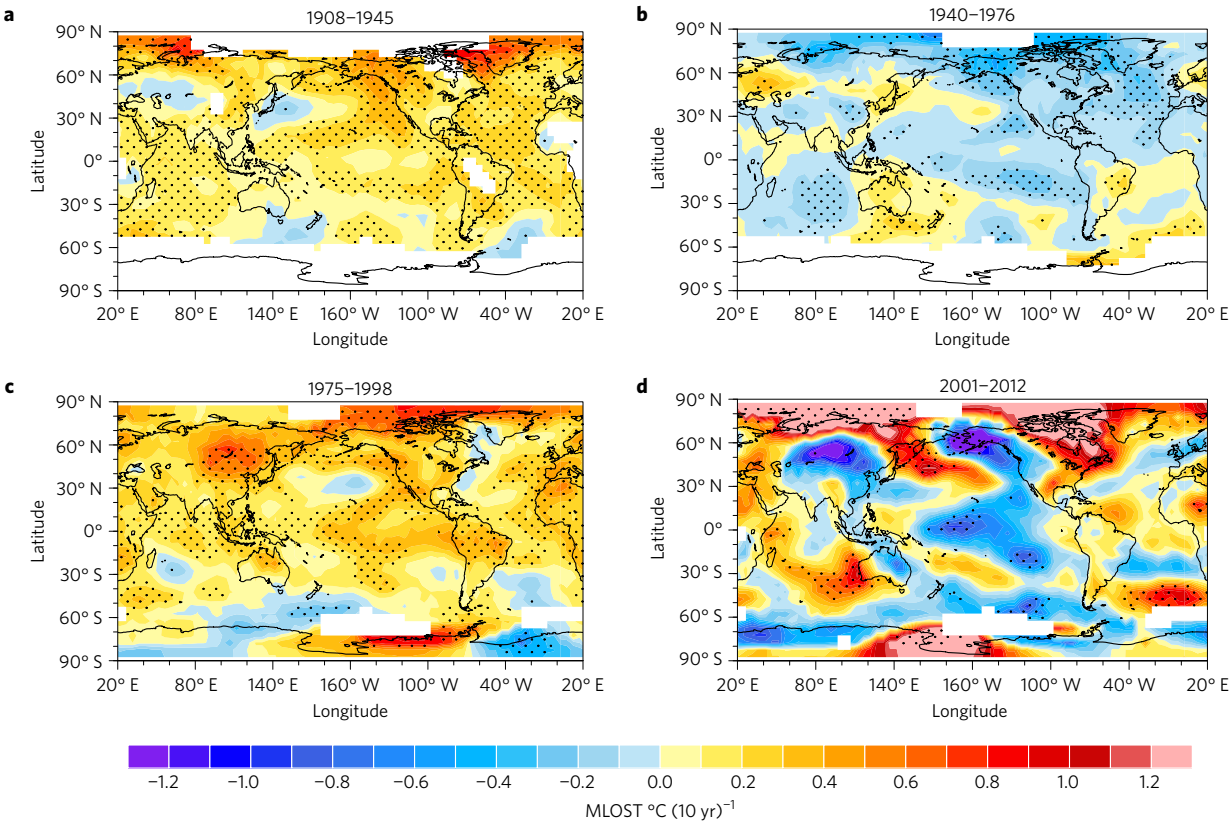



**Figure 2 | Observed annual-mean merged land air temperature and ocean surface temperature (MLOST) trend patterns during the two acceleration and slowdown periods. a**, 1908–1945. **b**, 1940–1976. **c**, 1975–1998. **d**, 2001–2012. The MLOST trends are estimated on the basis of the ensemble mean of four observations (Methods). White areas represent missing values. Stippling denotes 5% significance according to a two-sided Student's *t*-test.



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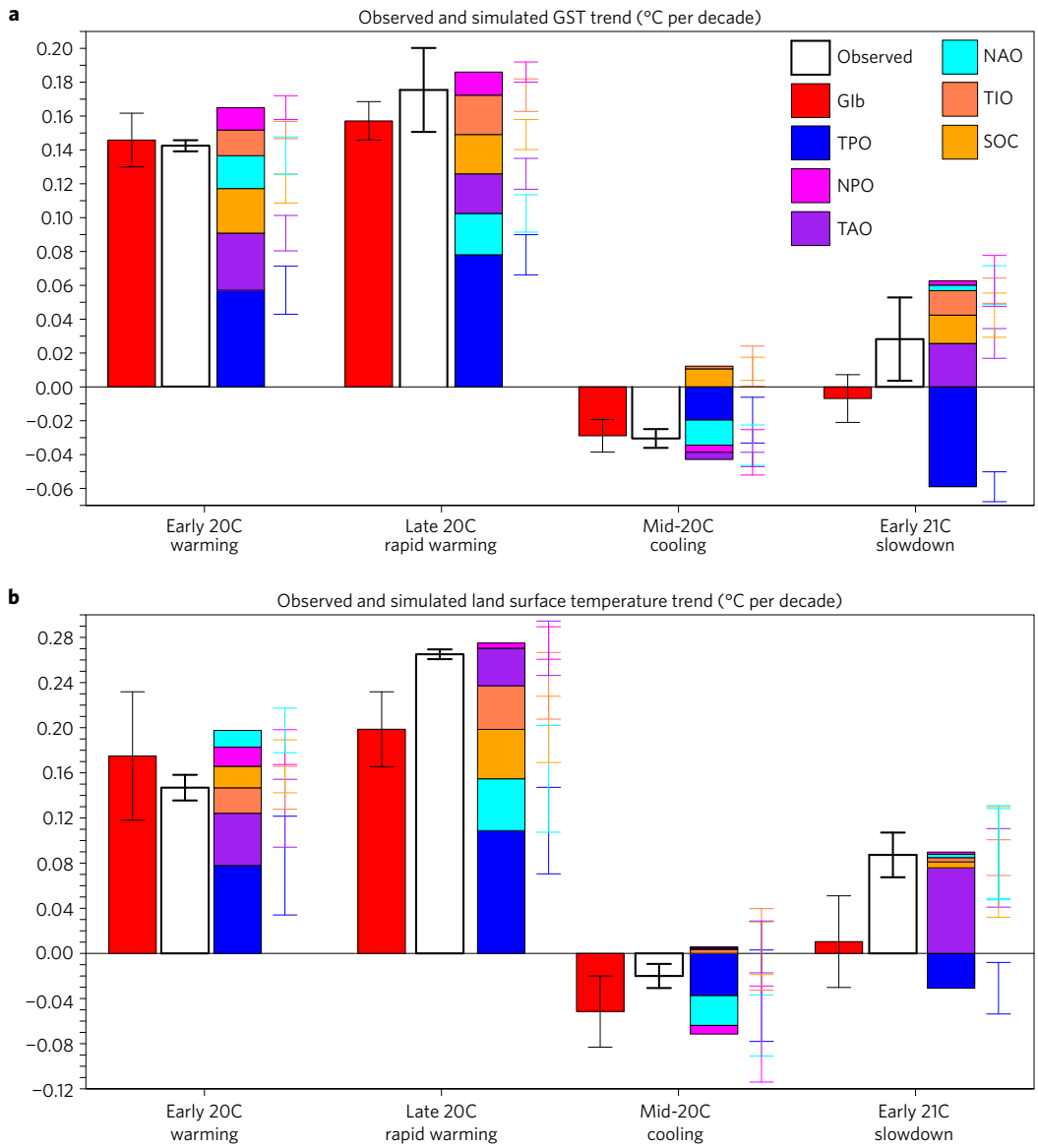



**Figure 3 | Near-surface temperature trends during the two acceleration and two slowdown periods. a**, Observed GST trends and simulated globally averaged near-surface air temperature trends. **b**, Observed and simulated land surface air temperature trends. The black empty bars indicate the observed ensemble mean GST and land surface air temperature of the four different data sources during each period. The simulated global near-surface air temperature trends include those forced by the global ocean (Glb), tropical Pacific Ocean (TPO), North Pacific Ocean (NPO), tropical Atlantic Ocean (TAO), North Atlantic Ocean (NAO), tropical Indian Ocean (TIO), and Southern Ocean (SOC). The model's simulations are based on six-member ensemble mean for each experiment (Methods). The order of the colour filled bars follows the order of the magnitudes of individual oceans' contributions. The error bars denote one standard deviation of the differences among the four observations (thick black bars) and across six members of each experiment (thin black bars for the global ocean forcing; and individual ocean forcings in colours indicated).

during the two acceleration periods. This is strongly at odds with the differences in GW rates between the slowdown and acceleration periods (compare Fig. 2 and Supplementary Table 1), although the amplified polar surface air temperature warming in the early twenty-first century (Fig. 2d) may partly offset the tropical Pacific-induced cooling effects[21].

A prominent feature during the two acceleration periods is that significant SST warming appears across almost the entire global ocean except patches of weak cooling in the extratropics (Fig. 2). Consistently, surface air temperature warming occurs over the majority parts of global continents (Fig. 2a,c). In contrast, SST cooling during the two slowdown periods does not extend to the entire global ocean (Fig. 2b,d). In particular, extensive warming appeared in the Atlantic, the Indian Ocean, and the extratropical Pacific during 2001–2012, opposite to the strong tropical Pacific cooling (Fig. 2d). Correspondingly, only limited areas of the global continents experienced significant cooling or warming during 2001–2012 (Fig. 2d). These results suggest that, while the tropical Pacific warming/cooling plays a major role, the SST changes in other basins may also have important influences on the GW rates[22–24].

As an attempt to quantify the relative contributions of individual ocean SST changes to the GW rates, we perform a set of atmospheric general circulation model experiments (Methods) in which the global SST trend during the two acceleration and slowdown periods is split into six regions, including the three tropical oceans that have predominant influences on the GST, and the North Atlantic, North Pacific and Southern Ocean (Supplementary Fig. 3). The simulations with the entire global SST trend forcing reproduce the



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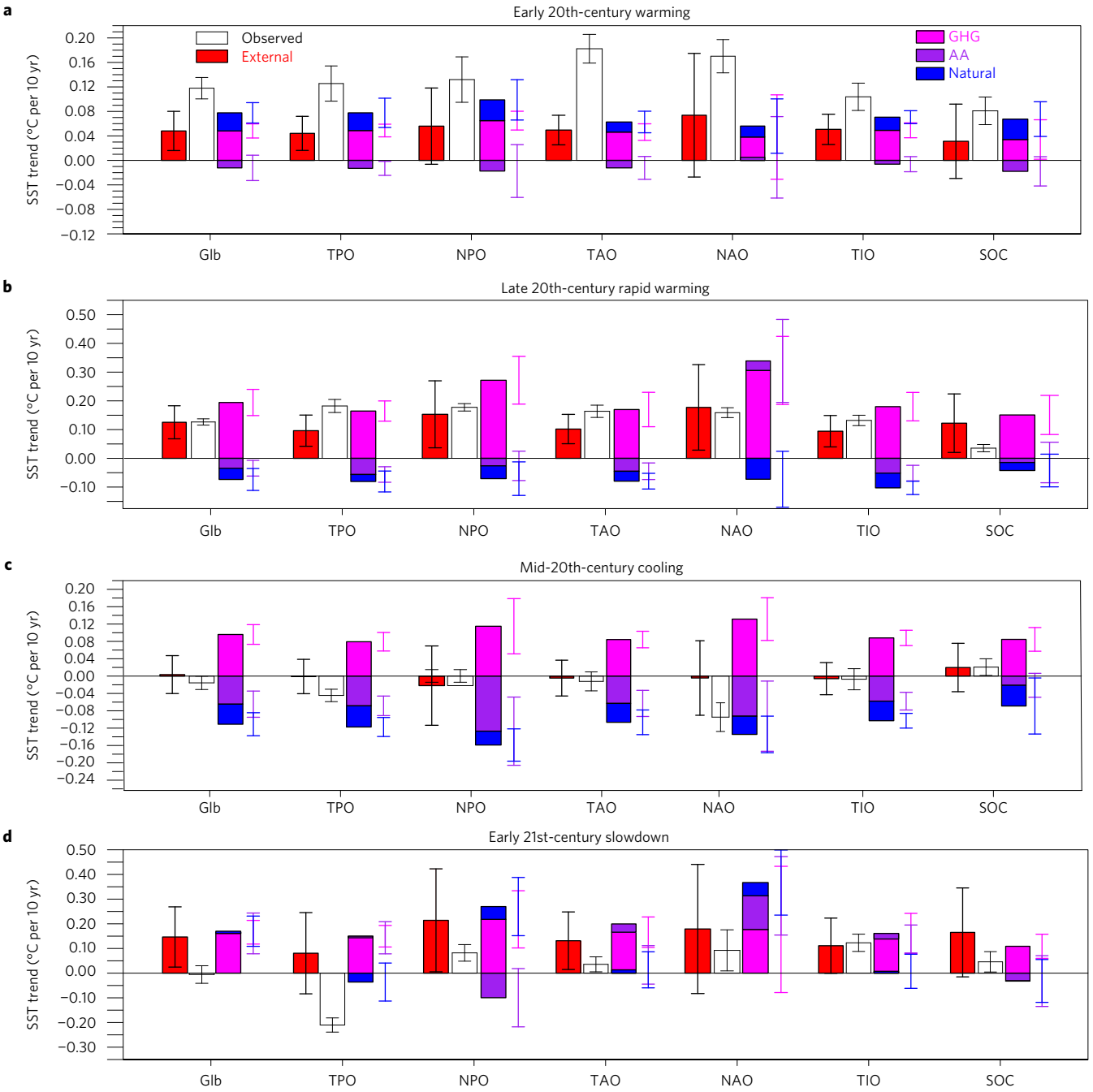



**Figure 4 | Observed and CMIP5-simulated SST trends in the global ocean and six individual ocean basins. a–d**, SST trends in the global ocean (Glb), tropical Pacific Ocean (TPO), North Pacific Ocean (NPO), tropical Atlantic Ocean (TAO), North Atlantic Ocean (NAO), tropical Indian Ocean (TIO) and Southern Ocean (SOC) from the observations, CMIP5 multi-model ensemble mean historical simulations (External, ideally represent the externally forced trends), multi-model ensemble mean simulation with greenhouse gas forcing only (GHG, Supplementary Table 2), with anthropogenic aerosol forcing only (AA), and with natural forcing only (Natural) for the early twentieth-century warming period (**a**), the late twentieth-century rapid warming period (**b**), the mid-twentieth-century cooling period ( **c**) and the early twenty-first-century slowdown period (**d**). The error bars indicate one standard deviation of the differences among four observations (thick black bars) and across CMIP5 models (thin black bars for historical simulations; and individual radiative forcings in colours indicated).

observed GST and land SAT warming during the early and late twentieth century fairly well (Fig. 3a,b, red filled and black empty bars). The model also successfully reproduces the observed GST and land SAT cooling in the mid-twentieth century (Fig. 3a,b). The recent slowdown, however, is not well simulated. While the observations show a muted warming, the model produces a weak GST cooling and a weak land SAT warming; this may be partly because Arctic amplification[21] is not captured (Supplementary Fig. 4d) due to the fact that observed sea-ice changes are not prescribed in the model. Besides, model's GHGs concentrations are fixed for each period and therefore the large direct forcing effect is not simulated (Methods).

The simulated land SAT trends are weaker than the observed (Supplementary Fig. 4c,d), particularly for the late twentieth-century rapid warming and recent slowdown periods; this suggests a role of direct external forcing[25] (Supplementary Information and Supplementary Table 3). The estimated direct forcing effect has



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persistently increased since 1900, in accordance with the increased trend of $CO_2$. Despite model deficiencies (Supplementary Information) and possible impacts of internal variability[26,27], the spatial distributions of observed GW trends during the two acceleration and slowdown periods (compare Fig. 2 and Supplementary Fig. 4) are generally well reproduced by the model with the global SST trend forcing, with pattern correlation reaching approximately 0.74, 0.82, 0.57 and 0.46, respectively.

The summed contributions of the SST changes in the six basins to the GW rates during the four inter-decadal periods are close (albeit not equal) to those simulated with the global SST trend forcing (Fig. 3, see Supplementary Information for the nonlinear effect discussion). During the two accelerated warming periods in the early and late twentieth century, the tropical Pacific (occupying ~21.1% of global area) SST warming plays the largest role, accounting for approximately 34% (39%) and 42% (39%) of the total GST (land SAT) warming rates, respectively (Fig. 3a,b). In the early twentieth century, the tropical Atlantic (~8.5% of global area) SST warming plays the second largest role (explaining ~20% and ~23% of the total GST and land SAT warming rates, respectively). In addition, for the two accelerated warming periods, the total SST warming in the Southern Ocean (~22.5% of global surface), North Atlantic (~4.1% of global area), tropical Indian Ocean (~8.4% of global area) and North Pacific (~5.8% of global area) accounts for nearly half and ~45% of the summed GST rates, respectively. These four oceans also explain large portions of land SAT warming rates (totally ~38% and ~49% during the two accelerated warming periods, respectively) and produce extensive warming over the global lands (Supplementary Fig. 5 and Supplementary Information).

During the mid-twentieth century, the cooling in the tropical Pacific plays an important role in the GST cooling, which is nearly equal to the summed contributions of the SST cooling in the North Atlantic, tropical Atlantic and North Pacific (Fig. 3a). Similarly, the summed contribution of the North Atlantic and North Pacific SST cooling to the global land SAT cooling is as important as that of the tropical Pacific (Fig. 3b). In contrast, the SST warming in the Southern Ocean and tropical Indian Ocean acts to increase the GST and land SAT by ~30% and ~7%, respectively, which tends to partly offset the cooling effects of the other oceans (Supplementary Fig. 6 and Supplementary Information). The compensation effect among the different oceans is more striking during the recent slowdown period. The strong cooling in the tropical Pacific generates a GST (land SAT) cooling of approximately −0.059 °C (−0.031 °C) per decade, consistent with the previous finding[5,6]. However, the induced GST and land SAT cooling is completely offset or overwhelmed by the summed warming effects of the tropical Atlantic and Indian Ocean, Southern Ocean, North Pacific and North Atlantic (Fig. 3 and Supplementary Fig. 7 and Supplementary Information).

In summary, our results reveal distinct impacts of individual basin's SST changes on the GW rates between the acceleration and slowdown periods. In the acceleration cases, the SST warming in all the oceans acts jointly to generate strong GW rates. In contrast, in the slowdown cases, the GST cooling induced by the tropical Pacific cooling is reduced or negated by the warming effects in the other oceans, leading to weak GW rates.

Thus, it is the SST changes in multiple oceans, not merely those in the tropical Pacific, that are crucial to determine the GW rates. On the basis of the Coupled Model Intercomparison Project Phase 5 (CMIP5) simulations with various radiative forcing (Methods), we explore possible causes of observed SST changes in individual oceans during the acceleration and slowdown periods. The observed SST warming in the global ocean and individual basins in the early twentieth century is partly reproduced by the multi-model historical simulations due to increased GHGs and natural forcing (Fig. 4a). Note that the difference between the observed and model simulations could be caused by internal variability[28] and errors in models[29] and historical forcings. In contrast, the late twentieth-century SST warming in the global ocean and most of the individual oceans is largely simulated in response to increased GHGs (Fig. 4b), which is partly offset by the cooling effects of anthropogenic aerosols and natural forcing.

In agreement with previous studies[30], our results suggest that increased anthropogenic aerosols during the mid-twentieth-century cooling period induce large cooling in global ocean and individual oceans, which overwhelms the warming effects of increased GHGs (Fig. 4c). However, large discrepancy is seen between the observed and multi-model historical simulations in the tropical Pacific and North Atlantic, which bears a resemblance to the observed SST trend pattern in the mid-twentieth century (Supplementary Fig. 8e). The discrepancy between the observations and multi-model historical simulations becomes more apparent in the tropical Pacific during the recent slowdown period (Fig. 4d and Supplementary Fig. 8g), suggesting that internal variability and/or model errors[29] may play an important role. The external radiative forcing produces generally uniform warming across the global ocean, largely accounting for the basin-wide average SST warming in the other oceans (Fig. 4d and Supplementary Fig. 8h).

The results imply that the SST changes in global and regional oceans during the two accelerations and slowdowns are driven by complicated and distinct mechanisms, which cannot be solely explained by modelled response to external radiative forcing. Improved understanding and projection of the distinct SST changes in multiple oceans between the multi-decadal GW acceleration and slowdown periods could help achieve a more accurate estimate of the future GW rate to better meet the GW target of the Paris Conference, compared with that based on increasing GHG emissions alone and/or a single-ocean SST change alone.

## Methods

Methods, including statements of data availability and any associated accession codes and references, are available in the online version of this paper.

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## Author contributions

S.-L.Y., J.-J.L. and G.H. conceived the research; J.-J.L. and S.-L.Y. designed the model experiments; S.-L.Y. and P.W. performed numerical simulations; S.-L.Y. performed the data analysis; J.-J.L. and S.-L.Y. wrote the paper; J.-J.L. and G.H. supervised this study.

## Additional information

Supplementary information is available in the online version of the paper. Reprints and permissions information is available online at www.nature.com/reprints. Publisher's note: Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations. Correspondence and requests for materials should be addressed to J.-J.L. or G.H.

## Competing financial interests

The authors declare no competing financial interests.



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## Methods

**Data sets.** We use the Hadley Centre-Climate Research Unit merged land air temperature and ocean surface temperature (MLOST) version 4.4.0.0 median (HadCRUT4)[31] and global land near-surface air temperature (SAT) version 4.4.0.0 (CRUTEM4)[32], the National Aeronautics and Space Administration Goddard Institute for Space Studies (NASA GISS) GISTEMP[33] with 1,200 km smoothing and global mean land SAT, the National Oceanic and Atmospheric Administration National Climatic Data Center (NOAA NCDC) MLOST version 4.0.0 (ref. 34) and the Global Historical Climatology Network land SAT version 3.3.0, the International Surface Temperature Initiative (ISTI) databank[14] MLOST time series and gridded data sets and land SAT time series for the period 1900–2014. We use monthly mean Centennial In Situ Observation-Based Estimates of the Variability of Sea Surface Temperature (COBE-SST) and version 2 (COBE-SST2)[35], corrected Extended Reconstructed Sea Surface Temperature version 4 (ERSST v4)[14], and merged Hadley-Optimal-Interpolation sea surface temperature and sea-ice concentration data sets[36] for the period 1900–2012. To reduce observational uncertainty, the ensemble mean of the observed MLOST, land SAT and SST from the four different data sets is employed in this study.

We also analyse the multi-model ensemble mean SST provided by the Coupled Model Intercomparison Project[37] (CMIP5) of the Intergovernmental Panel on Climate Change. 128 members of 42 models' historical simulations that are extended from 2006 to 2012 on the basis of RCP4.5 scenario runs, 35 members of 9 models' greenhouse gas (GHG) forcing simulations, 38 members of 10 models' natural forcing simulations, and 21 members of 5 models' anthropogenic aerosol forcing simulations for the period 1900–2012 (Supplementary Table 2) are analysed to investigate causes of the observed SST changes in multiple ocean basins (Fig. 4 and Supplementary Fig. 8). To reduce model uncertainty, we calculate the multi-model ensemble mean of each scenario experiment using the same weight for each model.

**Analysis methods.** We adopt an optimal piecewise linear regression to fit the trends with increasing time intervals of observed GST and MLOST. On the basis of the following three thresholds: a maximum magnitude of GST trends above 95% confidence level, relatively low observational uncertainty, and a high percentage of grid boxes that show statistical significant trends above 95% confidence level, we detect two robust accelerated warming periods in the early twentieth century (1908–1945) and the late twentieth century (1975–1998), and two global warming (GW) slowdown periods in the mid-twentieth century (1940–1976) and the early twenty-first century (2001–2012) (Supplementary Fig. 1). Note that some of the time windows are slightly overlapped, representing an uncertainty in defining the periods of GW accelerations and slowdowns. Removing the overlapped time intervals among these periods gives similar results. The least-square linear regression and Theil–Sen slope[38] methods are used to estimate the observed GST trends, with the statistical significance being performed with the Student's $t$-test and Mann–Kendall test (Supplementary Table 1). The observed SST and MLOST trends are computed at each grid box where the data availability covers more than 80% of the years during individual GW acceleration and slowdown periods.

**AGCM sensitivity experiments.** The Max Plank Institute for Meteorology atmospheric general circulation model[39] (AGCM, ECHAM5.4.1) is used to examine the response of globally averaged surface temperature (GST) to the observed SST trend forcing in the global ocean and individual oceans. Given the important impacts of the tropical SST on the GST, we first split the tropical oceans into the tropical Pacific Ocean (TPO), tropical Atlantic Ocean (TAO), and tropical Indian Ocean (TIO). The extratropical regions are then separated into the North Pacific Ocean (NPO), North Atlantic Ocean (NAO) and Southern Ocean (SOC) (Supplementary Fig. 3). We adopt ECHAM5 T63L31, which has a horizontal resolution of ~209 km and 31 vertical layers[39]. In the model sensitivity runs, we prescribe the observed SST trends in the global ocean and six different basins separately with carbon dioxide ($CO_2$) concentration being fixed to the level at the centre year of the period during 1908–1945, 1975–1998, 1940–1976 and 2001–2012, respectively. An additional model sensitivity experiment, in which the observed $CO_2$ trend during 1958–2012 is prescribed but the global SST forcing is fixed with the climatological monthly mean of 1981–2010, produces a GW rate of ~0.028 °C per decade and a land SAT warming rate of ~0.07 °C per decade that accounts for ~22% and ~32% of the observed GW rate of ~0.13 °C per decade and observed land SAT warming rate of ~ 0.22 °C per decade averaged during 1958–2012, respectively. To avoid potential instability, a linear buffer zone with a width of five latitudes and longitudes is applied to prescribe the observed SST trends in individual basins (for example, in the tropical Pacific Ocean, the linear buffer zone extends from 30° N to 35° N, 35° S to 30° S, and 115° E to 120° E). For each of the model experiments that are integrated for 30 model years, we perform six-member ensemble simulations in which the same SST and $CO_2$ forcing is prescribed but their initial conditions are different and provided by an additional model run with climatological SST forcing. Results shown are based on the six-member ensemble mean of individual model experiments. Note that the global mean results are similar if model outputs in areas where the observations are missing during the GW acceleration and slowdown periods are excluded in the model's calculations.

**Data availability.** The data that support the findings of this study are available from the corresponding authors on request. The model outputs from the ECHAM5 sensitivity experiments are available to readers on request. Other model data are available from the CMIP5 data portal (http://cmip-pcmdi.llnl.gov/cmip5/data_portal.html). Sources for the observed data sets are: HadCRUT4 (ref. 31) and CRUTEM4 (ref. 32) (accessed on 1 October 2015): http://www.metoffice.gov.uk/hadobs/index.html; GISTEMP[33] (accessed on 1 October 2015): https://data.giss.nasa.gov/gistemp; NOAA NCDC[34] (accessed on 30 August 2015): https://www.ncdc.noaa.gov/data-access/marineocean-data/mlost; ISTI[14] (accessed on 15 July 2015): http://www.surfacetemperatures.org; COBE-SST and COBE-SST2 (ref. 35) (accessed on 10 August 2015): https://climatedataguide.ucar.edu/climate-data/sst-data-cobe-centennial-situ-observation-based-estimates; ERSST v4 (ref. 14) (accessed on 10 August 2015): https://www.ncdc.noaa.gov/data-access/marineocean-data/extended-reconstructed-sea-surface-temperature-ersst-v4; Hurrell SST[36] (accessed on 15 August 2015): https://cdp.ucar.edu/security/loginout.htm.

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LETTERS

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# Patterns of the seasonal response of tropical rainfall to global warming

Ping Huang[1]*, Shang-Ping Xie[2,3,4]*, Kaiming Hu[1], Gang Huang[5] and Ronghui Huang[1]

**Tropical convection is an important factor in regional climate variability and change around the globe[1,2]. The response of regional precipitation to global warming is spatially variable, and state-of-the-art model projections suffer large uncertainties in the geographic distribution of precipitation changes[3–5]. Two views exist regarding tropical rainfall change: one predicts increased rainfall in presently rainy regions (wet-get-wetter)[6–8], and the other suggests increased rainfall where the rise in sea surface temperature exceeds the mean surface warming in the tropics (warmer-get-wetter)[9–12]. Here we analyse simulations with 18 models from the Coupled Model Intercomparison Project (CMIP5), and present a unifying view for seasonal rainfall change. We find that the pattern of ocean warming induces ascending atmospheric flow at the Equator and subsidence on the flanks, anchoring a band of annual mean rainfall increase near the Equator that reflects the warmer-get-wetter view. However, this climatological ascending motion marches back and forth across the Equator with the Sun, pumping moisture upwards from the boundary layer and causing seasonal rainfall anomalies to follow a wet-get-wetter pattern. The seasonal mean rainfall, which is the sum of the annual mean and seasonal anomalies, thus combines the wet-get-wetter and warmer-get-wetter trends. Given that precipitation climatology is well observed whereas the pattern of ocean surface warming is poorly constrained[13,14], our results suggest that projections of tropical seasonal mean rainfall are more reliable than the annual mean.**

The increase in atmospheric water vapour is a robust change under global warming as relative humidity remains relatively unchanged. The resultant intensification of the vertical moisture gradient, advected by the mean vertical motion, causes rainfall to increase where net water flux at the surface ($P$–$E$) is positive, and vice versa (wet-get-wetter)[7]. The enhanced lateral advection calls for reduced precipitation on the margins of a climatological rain band (upped-ante)[6,8]. The wet-get-wetter and upped-ante mechanisms do not emphasize the spatial variations in sea surface temperature (SST) warming. In Coupled Model Intercomparison Project (CMIP) models, the spatial distribution of tropical rainfall response is highly correlated with the overlooked SST pattern[5]. As tropospheric temperature is flattened by fast equatorial waves and set by the tropical mean SST (refs 10,15), local change in convective instability and hence rainfall is determined by relative SST defined as the deviation from the tropical mean (warmer-get-wetter, WaGW)[9,15]. We show that both the WaGW and wet-get-wetter (WeGW) mechanisms are important for tropical rainfall change, dominating the annual mean and seasonal anomalies, respectively. The result has implications for constraining rainfall projection.

The historical simulation and the representative concentration pathway 4.5 (RCP 4.5) experiment from 18 CMIP5 models (see Methods) are used to represent the present and future climates, respectively[16]. Figure 1 shows multi-model ensemble (MME) and zonal mean change in precipitation ($P$) and SST. The WeGW effect is obvious as $\Delta P$ moves back and forth across the Equator following the seasonal cycle of $\bar{P}$ (refs 17,18). (The overbar and $\Delta$ denote the present climatology and change in future climate, respectively.) The ocean warming ($\Delta$SST) peaks on the Equator and shows a weak seasonal cycle[9,19]. (Seasonal variations of $\Delta$SST are only 6% of the annual-mean, spatial variations.) On close inspection, the deviations from the WeGW pattern are apparent: precipitation increase ($\Delta P$) exhibits smaller meridional excursions than $\bar{P}$, with the maximum displaced on the equatorward flank of the climatological rain band. The peak rainfall is larger in March than September for $\Delta P$ but the opposite is true for $\bar{P}$. We will show that these deviations from WeGW are $\Delta$SST effects. (The upped-ante mechanism calls for rainfall reduction on both margins of the mean convergence zone, and does not explain the consistent equatorward displacement of $\Delta P$ relative to $\bar{P}$.)

To isolate the $\Delta$SST effect, we also analyse a pair of additional atmospheric experiments in CMIP5, forced respectively by a spatially uniform SST increase (SUSI) of 4 K and by the spatially patterned SST increase (SPSI) derived from MME CMIP3 quadruple $CO_2$ (1%to4×) simulations[16,20,21]. The observed SST climatology is used for both runs. The SST warming pattern used in the SPSI run is very similar to $\Delta$SST in RCP 4.5, both with an equatorial peak and a weak seasonal cycle (Supplementary Fig. S1). The SPSI results are scaled by its tropical mean SST increase, so the SPSI minus SUSI difference represents the SST pattern effect under the assumption that the nonlinear dependency on $\Delta$SST amplitude is small.

A band of increased precipitation in SUSI marches across the Equator following $\bar{P}$, and the peak of $\Delta P$ coincides with that of $\bar{P}$ (Fig. 2a). This validates the WeGW mechanism. Precipitation change in SPSI is similar to that in RCP 4.5 runs (Figs 2b and 1a) but deviates markedly from SUSI runs. Compared with SUSI, the anomalous rain band in SPSI features small meridional swings and rainfall increase is enhanced near the Equator. The SPSI minus SUSI difference shows only a weak seasonal cycle and is tightly trapped on the Equator, consistent with the $\Delta$SST structure. The $\Delta$SST effect is stronger in the first than in the second half of

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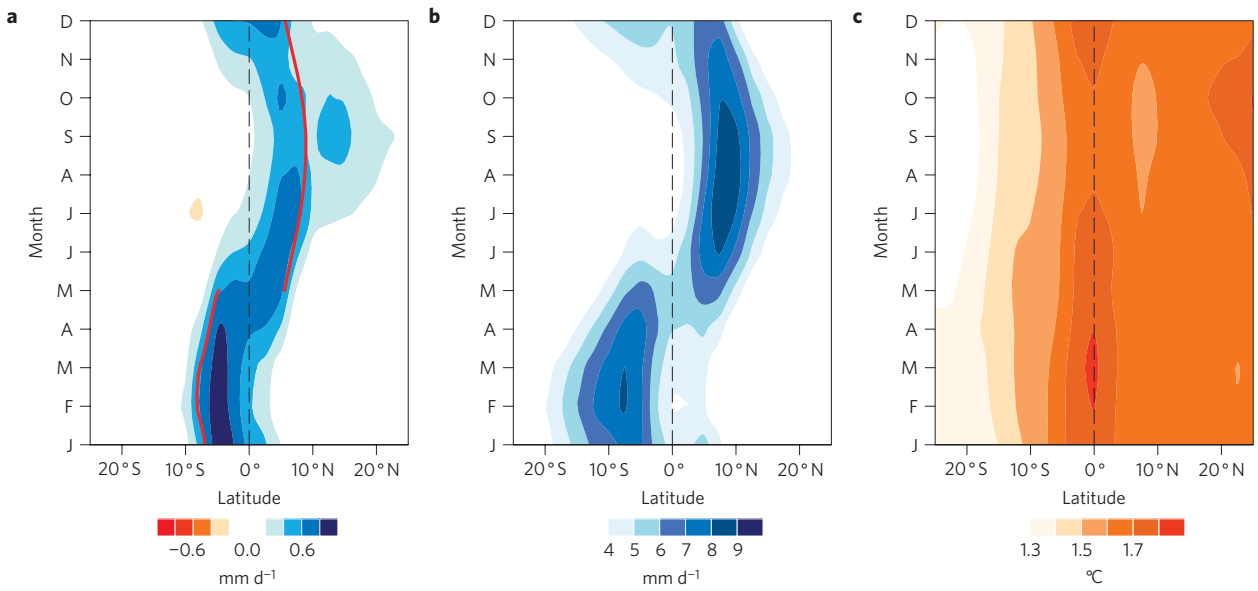



**Figure 1 | Seasonal cycle of precipitation and SST change. a–c**, Precipitation change (**a**), the precipitation climatology (**b**) and SST change (**c**) in RCP 4.5, all in zonal and MME mean. In **a**, the red curve marks the latitude of the maximum in mean precipitation.

the year because the seasonal development of equatorial upwelling suppresses the SST warming[9].

Vertical velocity change ($\Delta\omega$), overlooked in the WeGW hypothesis, proves important for rainfall change. In SUSI, $\Delta\omega$ broadly represents a weakening of tropical circulation (Fig. 2d) as required by a muted response of global precipitation[8,14]. In SPSI, $\Delta\omega$ is greatly enhanced near the Equator owing largely to the SST effect (Fig. 2e). The SST-induced change in $\Delta\omega$ closely resembles that in $\Delta P$ (Fig. 2f,c). The $\Delta$SST peak anchors the anomalous rising motion near the Equator and compensating subsidence on the flanks. Even the weak seasonal cycle is mutually consistent among $\Delta$SST, $\Delta\omega$ and $\Delta P$, modulated to some degree by convective feedback[22]. The time–latitude evolution of $\Delta\omega$ is similar between SPSI and RCP 4.5 runs (Figs 2e and 3a). The $\Delta$SST effect on $\Delta\omega$ can alternatively be understood from a moist instability argument[9]. $\Delta$SST and surface $\Delta q$ dominate the regional distribution of convective instability change because of the flattened upper tropospheric temperature change[9,10]. Specifically on the Equator where $\Delta$SST peaks, the enhanced convective instability anchors a maximum in precipitation increase and anomalous upward motion.

The water vapour budget can decompose tropical rainfall change into

$$\Delta P \sim \Delta\omega \cdot \bar{q} + \bar{\omega} \cdot \Delta q \qquad (1)$$

where $q$ denotes surface specific humidity, $\omega$ is the pressure velocity at 500 hPa, and lateral advection has been neglected, a good approximation for the tropics[7]. Figure 3 shows the decomposition results for RCP 4.5 runs. Equation (1) reproduces $\Delta P$ well from full models (Figs 1a and 3d). (The calculation using the full vertical-integrated budget yields nearly identical results.) As both $\bar{q}$ and $\Delta q$ have broad structures that peak in the climatological rain band (Supplementary Fig. S2), both terms in equation (1) are dominated by vertical velocity in meridional structure (Fig. 3), although the poleward moisture decrease reduces the effect of vertical motion on precipitation away from the deep tropics. The thermodynamic component $\bar{\omega} \cdot \Delta q$ represents the WeGW effect, with the upward motion in the climatological rain band pumping up the moisture increase near the surface. The dynamic component $\Delta\omega \cdot \bar{q}$ causes rainfall to increase (decrease) near (off) the Equator, consistent with the vertical circulation change induced by the equatorial peak in SST warming. We have repeated the decomposition for SUSI and SPSI runs (Supplementary Fig. S3). The results confirm that the dynamic component is due to the SST effect following the WaGW pattern. Remarkably, the thermodynamic component is nearly identical between SUSI and SPSI runs, illustrating that the seasonal cycle in precipitation change is dominated by the WeGW mechanism.

Monthly mean $\Delta P$ is significantly correlated with both $\bar{P}$ and $\Delta$SST at 0.67 and 0.60, respectively (Fig. 4), indicating that the WeGW and WaGW mechanisms are both important. The correlation is taken for the seasonal cycle within 10° S–10° N, a latitudinal band that pronounced precipitation change is confined to (Fig. 1a). To illustrate the combined WeGW–WaGW effect, we construct a multi-variant regression

$$\Delta P = a\bar{P} + b(\Delta\mathrm{SST} - c) \qquad (2)$$

for latitude–seasonal variations of zonal and MME means within 10° S and 10° N for 12 months. With $a = 0.11$, $b = 2.23$ and $c = 1.76$, the regression correlates with $\Delta P$ from CMIP5 models at 0.89, significantly higher than the correlation with either $\bar{P}$ or $\Delta$SST for the MME analysis. A similar multi-variant regression analysis is performed for each model. The coefficient $a \approx \beta \cdot \langle\Delta\mathrm{SST}\rangle$, where the angle brackets denote the tropical mean and $\beta = 0.07\,\mathrm{K}^{-1}$ is the Clausius–Clapeyron coefficient, represents the thermodynamic effect on $\Delta P$, whereas the coefficient $c \approx \langle\Delta\mathrm{SST}\rangle$ relates the change in the SST threshold for atmospheric convection to the tropical mean SST warming[10] (Supplementary Fig. S4). We use a non-dimensional ratio $\alpha = (b \cdot \sigma(\Delta\mathrm{SST} - c)/a \cdot \sigma(\bar{P}))$ to measure the relative importance of the WaGW to WeGW effect, where $\sigma$ denotes the standard deviation. For MME, $\alpha = 0.9$, indicating that the WeGW and WaGW effects contribute nearly equally to $\Delta P$. MME average generally underestimates the SST pattern and its effect on rainfall. Indeed the same regression analysis for individual models yields $\alpha > 0.9$ for 13 out of 18 models (Fig. 4d). (Results are not shown for FGOALS-s2 and MIROC5, in which the $\Delta$SST pattern effect is so strong that $\Delta P$ is negatively correlated with $\bar{P}$.) The ratio $\alpha$ is significantly correlated with the standard deviation of SST pattern at 0.55 among models, indicating that the stronger the SST pattern is, the greater its relative importance for precipitation change.

In summary, we show that the seasonal precipitation response to global warming is a hybrid of WeGW and WaGW effects (WeWa).

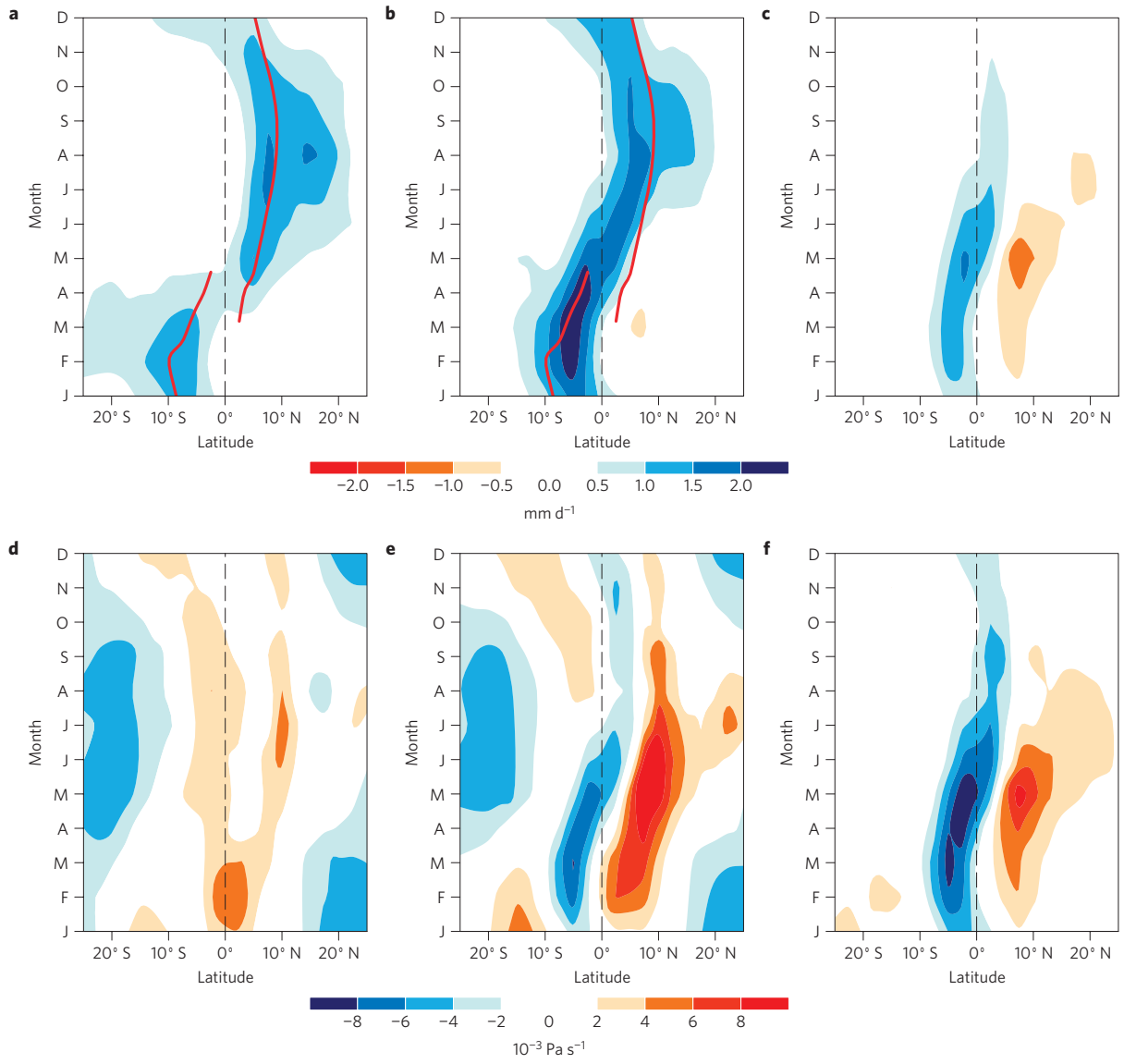



**Figure 2 | Seasonal cycle of precipitation and circulation changes in SUSI and SPSI runs. a–c**, $\Delta P$ in SUSI (**a**) and SPSI (**b**) runs, and their difference (**c**). **d–f**, The same as in **a–c** but for pressure velocity $\Delta\omega$ at 500 hPa. The solid red lines in **a**,**b** mark the latitude of the climatological precipitation maximum in the control. The SPSI results are scaled by the 20° S–20° N mean ΔSST to match SUSI.

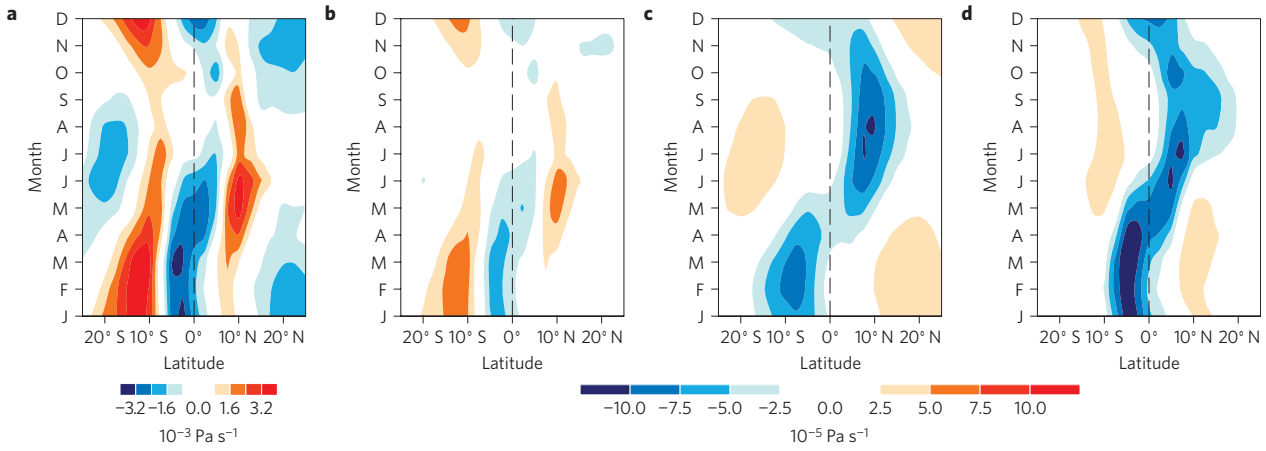



**Figure 3 | Circulation change and decomposition of precipitation change. a–d**, $\Delta\omega$ at 500 hPa (**a**); dynamic ($\Delta\omega \cdot \bar{q}$) (**b**) and thermodynamic ($\bar{\omega} \cdot \Delta q$) (**c**) components of rainfall change and their sum (**d**).

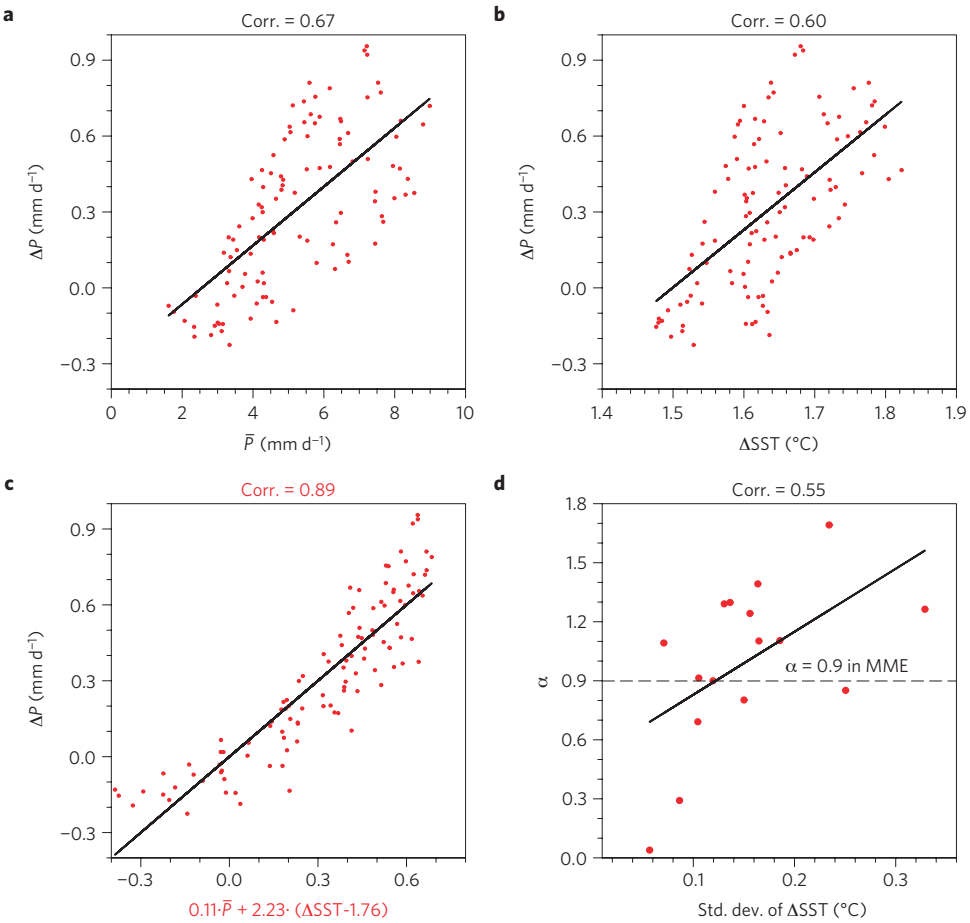



**Figure 4 | Relationship of precipitation change to mean precipitation, SST change pattern and their linear combination. a–c**, Scatter plots of MME precipitation change with mean precipitation (**a**), SST change (**b**) and their linear combination (equation (2); **c**) in 10° S–10° N through the seasonal cycle. The solid line indicates the linear fit, with the correlation coefficient shown at the top. **d**, The multi-variant regression analysis is also performed for each model, and the ratio α is shown in relation to the spatial standard deviation of ΔSST.

The two effects are not mutually exclusive but complementary, and the combined WeWa view best explains the seasonal precipitation change in the tropics (Fig. 4c). The anomalous rain band moves back and forth across the Equator but is consistently displaced on the equatorward flank of the climatological rain band. The displacement is due to the SST pattern effect. Specifically, the peak of SST warming anchors a band of anomalous ascent and rainfall increase near the Equator. Superimposed on this annual mean pattern, the upward motion in the climatological convergence zone sucks up the moisture increase from the boundary layer like a vacuum cleaner, dragging the band of rainfall increase back and forth across the Equator. We have repeated the linear regression analysis for the annual mean, and two seasons when the climatological rain band is farthest away from the Equator (February–April and August–October). The ratio of the WaGW to WeGW effect α is 1.35, 0.71 and 0.63, respectively (Supplementary Fig. S5). Thus, the WaGW effect is more obvious in the annual mean rainfall change whereas the WeGW effect is more dominant for the seasonal mean.

Regional precipitation projection is important for adaptation but subject to inherent uncertainty much greater than temperature[23,24]. The WeGW and WaGW effects are distinct in the nature of uncertainty regarding regional projection. If the WeGW mechanism dominates, the well-observed climatology can be used to constrain rainfall projection[25]. If the WaGW mechanism were to dominate, on the other hand, the uncertainty would be much greater because the SST warming pattern varies among models[5,26] and is poorly constrained by observations[13,14,27]. In light of the fact that the WeGW effect contributes more to the seasonal than the annual mean rainfall, we suggest that model projection is more reliable for the seasonal than annual mean rainfall in the tropics. This explains why the global monsoon rainfall increases consistently across models[28,29], a change that our results suggest probably takes place, with great environmental and socio-economic impacts.

## Methods

**CMIP5 models.** We use the historical and RCP 4.5 experiments from 18 coupled general circulation models contributing to CMIP5 at http://pcmdi9.llnl.gov/ (ref. 16). They are BCC-CSM1.1, CanESM2, CCSM4, CNRM-CM5, CSIRO-Mk3.6.0, FGOALS-s2, GFDL-CM3, GFDL-ESM2G, GISS-E2-R, HadGEM2-ES, INM-CM4, IPSL-CM5A-LR, IPSL-CM5A-MR, MIROC5, MIROC-ESM, MIROC-ESM-CHEM, MRI-CGCM3 and NorESM1. See http://cmip-pcmdi.llnl.gov/cmip5/availability.html for details. The 1981–2000 mean defines the present climatology, the 2081–2100 mean in RCP 4.5 runs the future climatology, and their difference represents the change under global warming. The MME is defined as the simple average of 18 models.

Output from six atmospheric models (CanAM4, CNRM-CM5, HadGEM2-A, IPSL-CM5A-LR, MIROC5 and MRI-CGCM3) is available for a set of control, SUSI, and SPSI runs, called amip, amip4K and amipFuture in CMIP5, respectively[16,21]. In SUSI, a spatially uniform SST increase of 4 K (constant in time) is superimposed on the observed SST whereas SPSI uses the spatially patterned SST increase derived from the MME CMIP3 quadruple $CO_2$ (1%to4×) simulation. The SST pattern in SPSI is similar to that in CMIP5 RCP 4.5 runs, both with a peak on the Equator and a weak seasonal cycle (Supplementary Fig. S1 and Fig. 1c). The climate change

is defined as the 20-year monthly climatology for 1981–2000 in SUSI and SPSI runs minus that in the control runs. We scaled the SPSI results so that the tropical mean SST increase is the same as SUSI.

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## Author contributions

P.H. designed and performed the analysis. S-P.X. contributed to improving the analysis and interpretation. K.H. and G.H. prepared part of the data. P.H. and S-P.X. wrote the paper. All authors discussed and commented on the paper.

## Additional information

Supplementary information is available in the online version of the paper. Reprints and permissions information is available online at www.nature.com/reprints. Correspondence and requests for materials should be addressed to P.H. or S-P.X.

## Competing financial interests

The authors declare no competing financial interests.

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# Interdecadal modulation of El Niño amplitude during the past millennium

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**The El Niño/Southern Oscillation (ENSO) is the dominant mode of interannual climate variability on Earth, alternating between anomalously warm (El Niño) and cold (La Niña) conditions in the tropical Pacific at intervals of 2–8 years[1,2]. The amplitude of ENSO variability affects the occurrence and predictability of climate extremes around the world[3,4], but our ability to detect and predict changes in ENSO amplitude is limited by the fact that the instrumental record is too short to characterize its natural variability[5–8]. Here we use the North American Drought Atlas[9,10]—a database of drought reconstructions based on tree-ring records—to produce a continuous, annually resolved record of ENSO variability over the past 1,100 years. Our record is in broad agreement with independent, ENSO-sensitive proxy records in the Pacific and surrounding regions. Together, these records indicate that ENSO amplitude exhibits a quasi-regular cycle of 50–90 years that is closely coupled to the tropical Pacific mean state. Anomalously warm conditions in the eastern Pacific are associated with enhanced ENSO variability, consistent with model simulations[11]. The quasi-periodic ENSO amplitude modulation reported here offers a key observational constraint for improving models and their prediction of ENSO behaviour linked to global warming.**

Despite great efforts to understand the El Niño/Southern Oscillation (ENSO) phenomenon, there is still much uncertainty in changes in ENSO amplitude. Instrumental records indicate marked variations in ENSO amplitude, large around the 1890s–1900s and 1970s–1990s, but small between these periods and before 1890 (Supplementary Fig. S1). Existing proxy records suggest significant variability in ENSO amplitude in the past[2,12,13], but they cover only the past few centuries or noncontinuous periods of the past millennium. Current general circulation models (GCMs) give different projections of future ENSO-related change, with some showing an increase and others no change or even a decrease in ENSO amplitude[6,7]. Therefore, it is crucial to develop long, continuous, high-resolution proxy records to determine the characteristics of natural ENSO variability, and to provide observational constraints for improving the models[7,8].

Proxy records from the equatorial Pacific are valuable for studying past ENSO behaviour[12–14], yet very few such records exist. This situation has impeded the study of long-term ENSO amplitude modulation. By contrast, thousands of tree-ring records without these deficiencies are available for North America, a region where hydroclimate is sensitive to tropical Pacific climate anomalies on various timescales[15,16]. Such tree-ring records have been used to develop the gridded North America Drought Atlas[9,10] (NADA) (see Methods and Supplementary Information). Here we demonstrate the fidelity of NADA in recording long-term ENSO variability, particularly variations in ENSO amplitude on interdecadal to centennial timescales.

The leading empirical orthogonal function (EOF) of NADA, which accounts for 23.5% of the total variance, shows a distinct moisture pattern heavily loaded over southwest North America (Methods and Fig. 1a), a region strongly affected by ENSO variability[15,16]. The explained variance of the first EOF for the southwest North America domain rises to 49.5% (Supplementary Fig. S3), and its principal component (PC) has a correlation of 0.99 with that for the whole of North America over the past 1,100 years. This confirms that NADA PC1 dominates variability in southwest North America.

The fidelity of NADA PC1 in representing ENSO variability is demonstrated in five ways: (1) NADA PC1 is significantly correlated with equatorial Pacific sea surface temperatures[17] (SSTs) during the instrumental period (Fig. 1c). For example, its correlation with the January–March (JFM) ENSO Niño3 index for 1870–2002 is 0.51 ($P < 0.001$) (Supplementary Fig. S4). (2) NADA PC1 is significantly correlated with a modern coral record from Palmyra Island in the central tropical Pacific[13], at $r = -0.58$ for the common period 1891–1994 ($P < 0.001$) (Supplementary Fig. S5). After adjusting relict coral U/Th dates within the analytical error windows[13], we found possible correlations between North American drought and tropical corals that persisted throughout the past millennium (Methods and Fig. 1d). Considering the independent nature of these proxy records, this agreement is remarkable, strongly indicative of a coherent relationship between ENSO and North American drought over the past millennium. (3) NADA PC1 is highly correlated with, but is significantly longer than, the existing ENSO reconstructions (Supplementary Table S1). Most of these reconstructions are largely, although not entirely, independent, as they only share a few common tree-ring data from southwest North America. This suggests that NADA PC1 represents reasonable estimates of ENSO variability over the past millennium. (4) Separate composite analysis of NADA for periods of large and small PC variability over the past millennium shows nearly identical patterns (Supplementary Fig. S6). This pattern stability suggests that the westerly waveguide is stable in tropical forced stationary waves, and that NADA PC1 represents the modulation of ENSO itself. (5) Separate EOF analysis shows that the first EOF of NADA is heavily loaded over southwest North America both before and after the 1976/1977 climate regime shift[18,19], and the PC is highly correlated with tropical eastern Pacific SSTs for both periods

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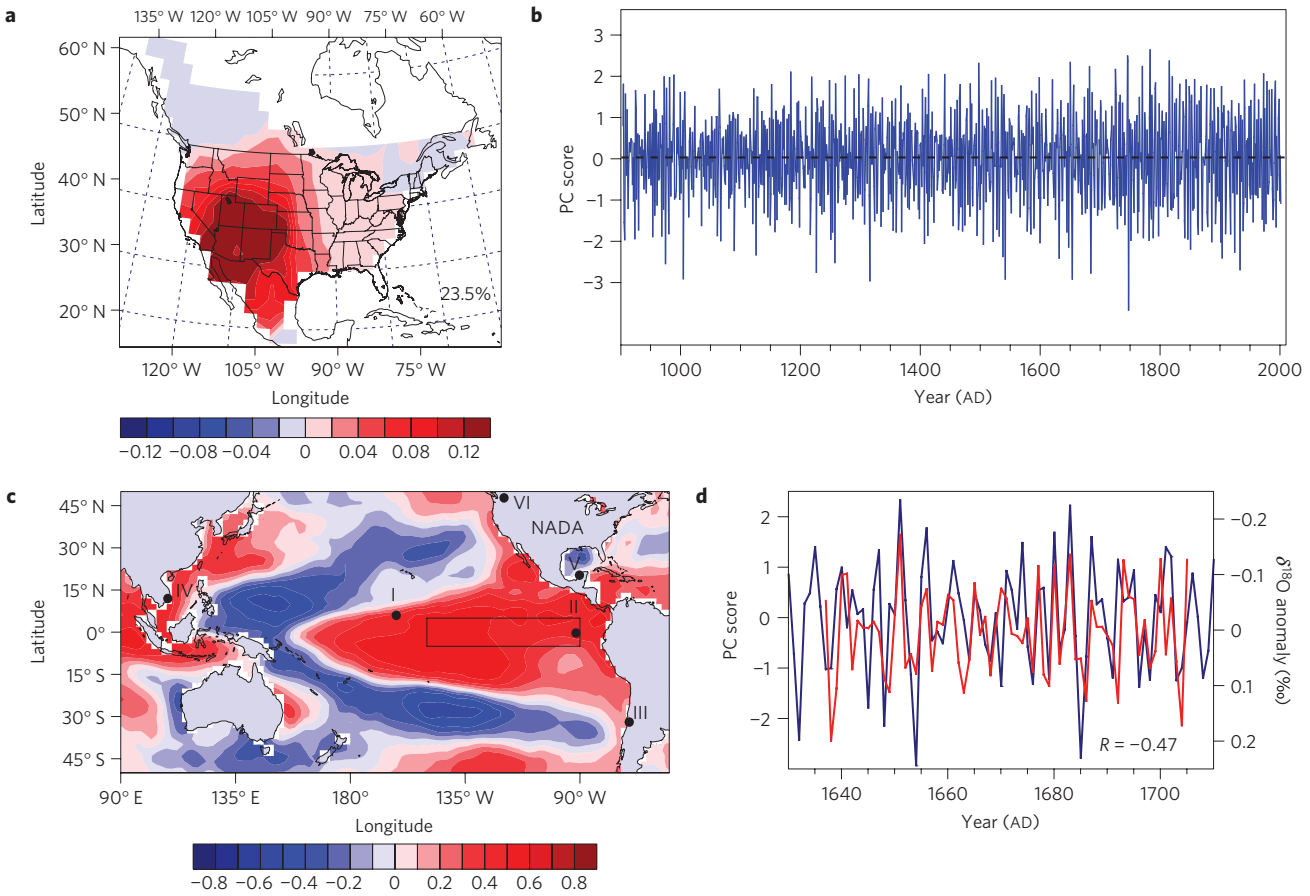



**Figure 1 | The leading EOF pattern of NADA, its PC series and correlation of the PC series with tropical records. a**, Spatial pattern of the leading EOF over North America, and **b**, its PC time series. The percentage variance accounted for by the leading EOF is labelled in **a**. **c**, Spatial correlation of NADA PC1 with global JFM SSTs (ref. 17) during 1870–2002. The SSTs were processed using a 9-yr Lanczos high-pass filter[30] to isolate interannual variability. The box denotes the Niño3 region, and the dots indicate the site locations mentioned in the text (I: Palmyra corals[13], II: Galápagos coral[12], III: Chile tree rings[23], IV: Vietnam tree rings[24], V: Mexico sediment[26], VI: Washington sediment[27]). **d**, Comparison of NADA PC1 (blue) with the dating-error corrected Palmyra relict coral $\delta^{18}$O series[13] (red) over 1637–1705.

(Supplementary Fig. S7). Although there exist some differences in the EOF pattern and correlation with tropical SSTs between the two periods, these results indicate that moisture variability in southwest North America has been sensitive to ENSO variability both before and after the 1976/1977 climate regime shift. The relationship to ENSO is further confirmed by the SST correlation pattern for 1870–2002 (Fig. 1c). Therefore, teleconnections of canonical eastern Pacific ENSO (hereafter simply ENSO) are probably stable over southwest North America. On the basis of these validations, we consider that NADA PC1 represents a continuous record of interannual ENSO variability during the past millennium.

As inferred from NADA PC1, ENSO amplitude has been highly variable during the past millennium, with many El Niño/La Niña events of larger amplitude than observed in the instrumental period (Fig. 1b). Here we calculated a 21-yr running biweight variance[20] to measure changes in ENSO amplitude. This technique highlights the interdecadal amplitude modulation of a time series while reducing bias that might be introduced by extreme outliers in a 21-yr window. As shown in Fig. 2a, superimposed on a general rising trend are large fluctuations in ENSO variance on interdecadal to centennial timescales. ENSO variance was reduced in the Medieval Climate Anomaly (MCA) period, with the eleventh-century variance being the lowest of the past millennium. Following this, the variance showed an increasing trend in the Little Ice Age (LIA), and maintained a high level from the eighteenth century to the Current Warm Period (CWP). On average, the eighteenth-century ENSO variance was the strongest of the past millennium.

A multi-taper method[21] (MTM) spectral analysis reveals two significant quasi-regular periodicities at 50–60-yr and 82–90-yr bands, respectively (Fig. 3a), confirming the visual impression of interdecadal modulation of ENSO amplitude. A time-space wavelet transform[22] exhibits a marked shift in the dominant periodicity, concomitant with the transition from the MCA into the LIA, from a 82–90-yr cycle during AD 900–1300 to a 50–60-yr cycle thereafter (Fig. 3b). Meanwhile, a 30-yr cycle is present during AD 1500–1800 and in the MTM analysis. The shifts in ENSO periodicity, together with a general increase in variance (Fig. 2a), may reflect major reorganizations within the tropical ocean–atmosphere system, or alternatively significant ENSO teleconnection changes from the MCA to LIA.

We found evidence for such interdecadal fluctuations in ENSO variance elsewhere, especially in the equatorial Pacific and regions under strong ENSO teleconnections. Variance modulation in North American tree rings is consistent with that observed in the Niño3 SST, and the two variance series are correlated at 0.64 ($P < 0.001$) during their common period 1880–1992 (Fig. 2a). A coral annual average oxygen isotopic ($\delta^{18}$O) record from the Galápagos Islands in the eastern equatorial Pacific[12] exhibits quasi-periodic variance modulation similar to that in North American tree rings during the past four centuries (Fig. 2b). The coral $\delta^{18}$O records from Palmyra Island in the central tropical Pacific[13], although not continuous,

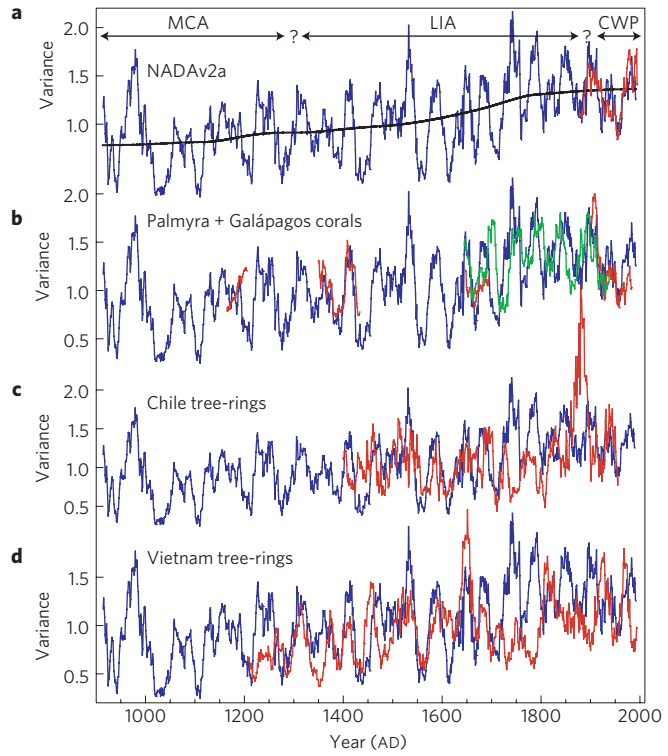



**Figure 2 | Comparison of records of ENSO variance.** NADA-derived ENSO variance (blue) along with those derived from: **a**, the observed Niño3 SST (ref. 17; red) over 1880–1995; **b**, coral records from both the Galápagos Islands in the eastern equatorial Pacific[12] (green) and Palmyra Island in the central tropical Pacific[13] (red); **c**, tree-ring record from an ENSO-sensitive site in El Asiento, Chile[23] (red); **d**, tree-ring record from the Bidoup Nui Ba National Park in southern Vietnam[24] (red). All records were processed using a 9-yr Lanczos high-pass filter[30] before calculating a 21-yr running biweight variance. The black bold line in **a** denotes a 400-year low-pass filter to indicate the long-term trend.

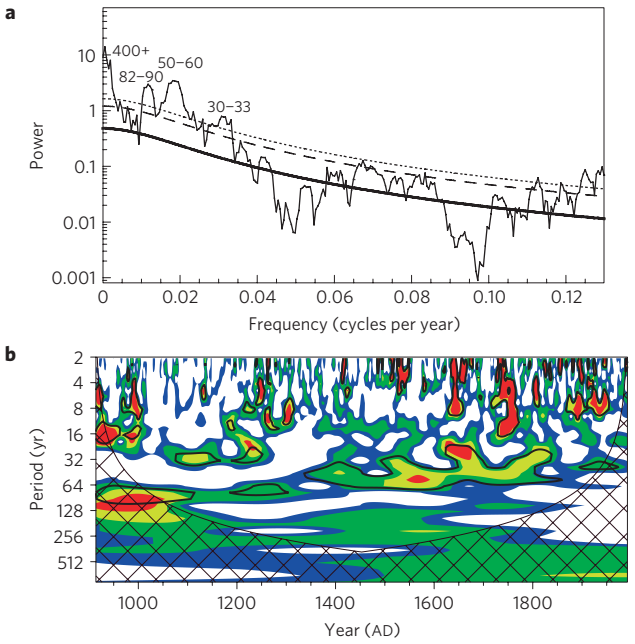



**Figure 3 | Spectral property of the NADA-derived ENSO variance series.** **a**, MTM (ref. 21) spectral density of the ENSO variance series. The bold line indicates the null hypothesis, and the dash (dotted) line indicates the 95% (99%) significance level, using a red-noise background spectrum. Periodicities of significant peaks are labelled above the spectral density line. **b**, The wavelet power spectrum[22]. The power has been scaled by the global wavelet spectrum. The cross-hatched region is the cone of influence, where zero padding has reduced the variance. Black contour is the 90% significance level, using a red-noise background spectrum.

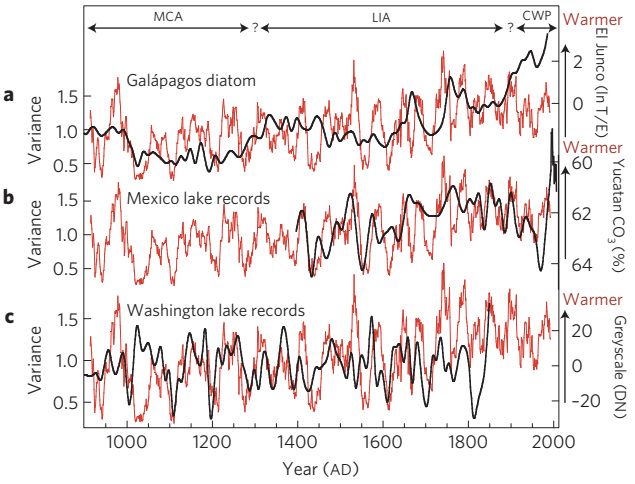



**Figure 4 | Comparison of the NADA-derived ENSO variance series with records of the tropical Pacific mean state.** NADA-derived ENSO variance (red) along with: **a**, a diatom record from El Junco Lake, Galápagos, in the eastern equatorial Pacific[14] (black). **b**, A carbonate ion ($CO_3^{2-}$) record from Aguada X'caamal Lake in northwest Yucatan, Mexico[26] (black). **c**, A greyscale record from Castor Lake in Washington state, USA (ref. 27) (black). All lake records were smoothed using a 9-yr low-pass filter. Upward swings in black curves represent a warm state of the tropical Pacific.

show broadly consistent changes in variance during the past nine centuries (Fig. 2b). Tree-ring records from an ENSO-sensitive site in El Asiento, Chile, in the Southern Hemisphere[23], exhibit similar interdecadal fluctuations in variance over most of the past six centuries, except during the late nineteenth century, when there is an out-of-phase relationship (Fig. 2c). Recent tree-ring records from the highlands of Bidoup Nui Ba National Park in southern Vietnam[24], a tropical region strongly influenced by ENSO teleconnections, also show in-phase fluctuations in variance with North American tree-rings during the thirteenth to seventeenth and most of the twentieth centuries, although this relationship was out-of-phase during the eighteenth to nineteenth centuries (Fig. 2d). We note two caveats of the above proxy comparisons. First, unlike the NADA, which is based on nearly two-thousand tree-ring chronologies across North America, other proxy data often consist of only a single series for a given site, and are prone to chaotic noise and regional variability. Second, the long-term stability of ENSO influence in other teleconnection regions needs further investigation, unlike North America where we have validated a persistent relationship to ENSO during the past millennium. Overall these records from the tropics and mid-latitudes in both the Northern and Southern Hemispheres are mutually consistent. Together they show quasi-periodic amplitude modulation, and suggest that such modulation is fundamental to ENSO physics.

How ENSO amplitude modulation is related to changes in the tropical Pacific mean state is a key question for ENSO physics[6,11,25]. Here we examine this relationship during the past millennium by comparing the NADA-derived ENSO variance series with three proxy records for the tropical Pacific mean state. The first is a diatom record from El Junco Lake, Galápagos, an indicator of SSTs in the eastern equatorial Pacific[14] (Fig. 4a). The second is a carbonate ion ($CO_3^{2-}$) record from Aguada X'caamal Lake in

northwest Yucatan, Mexico (Fig. 4b) that indicates SST changes in the surrounding ocean[26]. The third is a greyscale record from a sediment core from Castor Lake in the Pacific Northwest, whose low-frequency variations are interpreted as changes in the tropical and extratropical Pacific mean state[27] (Fig. 4c).

Despite discrepancies that may reflect either data uncertainties or changes in SST spatial structure, these three records indicate that the eastern tropical Pacific was in a cooler state at the peak of the MCA (~AD 1000–1200), and that there was a persistent positive temperature trend in the LIA and CWP (Fig. 4). ENSO variance was thus greatly reduced during most of the MCA, increased persistently in the LIA, and maintained a high level in the CWP. Moreover, interdecadal fluctuations in ENSO variance seem closely related to anomalies in the Pacific mean state during the past millennium. The Castor Lake data, in particular, suggest that a warm (cool) eastern tropical Pacific is associated with increased (reduced) ENSO variance over most of the 1,000-year record (Fig. 4c). These observations suggest that ENSO variability during the past millennium is closely coupled to changes in the background mean state, a finding that is supported by recent model simulations[6,11].

The quasi-regular nature of ENSO amplitude modulation, and the co-variations with the equatorial Pacific mean state, support the emerging idea that there is a positive feedback between the mean state and ENSO variance[11]. An increase in ENSO amplitude causes the eastern equatorial Pacific to warm, a mean-state change that amplifies ENSO variability. If fully tested in observations and process-oriented model studies, this idea can be exploited for decadal prediction of ENSO amplitude.

In summary, using data on North American tree rings, we have produced a continuous, annually resolved record of ENSO variability for the past millennium. This record is in broad agreement with independent proxy records in and surrounding the Pacific over most of their common periods, revealing a marked quasi-regular interdecadal modulation of ENSO amplitude throughout the past millennium. The amplitude modulation is closely associated with changes in the tropical Pacific mean state, suggesting an interactive relationship. This observation alone implies that the continued increase in eastern-central tropical Pacific SSTs in a future warmer climate[5,7] will lead to enhanced ENSO variability and more extreme climate conditions around the globe. The diverse response of a host of ocean–atmospheric feedbacks complicates such a prediction of ENSO amplitude change[5,7]. Regardless, the quasi-periodic modulation of ENSO amplitude provides a crucial observational constraint that can be used to guide model improvement towards reduced uncertainties in predicting future ENSO behaviour. Finally, the further study of the mechanisms responsible for such ENSO amplitude modulation (for example, stochastic atmospheric noise[28] and external forcing[29]) will help shed light on ENSO dynamics.

## Methods

**The updated North America Drought Atlas (NADA).** The NADA (ref. 9) consists of annually resolved summer (June–August) Palmer Drought Severity Index (PDSI) reconstructions derived from tree rings, available on a 2.5° × 2.5° grid over most of North America (Supplementary Information). Compared with the initial release of the NADA, the updated version (NADAv2a) used more tree-ring chronologies (1845 in total) to improve the PDSI reconstructions, but the grid spacing, series length, and statistical methods remain unchanged[10]. Here we analyse the updated NADA for AD 900–2006, a period when reconstructed values are available at more than 60% of all grid points (Supplementary Fig. S2). Time series at each grid point were processed using a 9-yr Lanczos high-pass filter[30] to preserve variability in the classical ENSO band of 2–8 years[1,2]. Empirical orthogonal function (EOF) analysis was used to objectively define the dominant mode of summer drought over NA, and the associated principal component (PC) series was used to indicate its variability over time.

**Relict coral dating error corrections.** By splicing together relict coral oxygen isotopic ($\delta^{18}$O) records from Palmyra Island in the central tropical Pacific, Cobb *et al.*[13] provided 30-150-yr windows of tropical Pacific climate variability over the past 1,100 years. Their modern coral samples were exactly dated, whereas relict corals were dated using the U/Th method, which contains a general dating error of ± 5–10-yr. Therefore, if there were a coherent relationship between ENSO and North American drought over the past millennium, then, by adjusting the relict coral sequences within the range of their U/Th-dating errors, a firm match could be achieved between the coral $\delta^{18}$O sequences and NADA PC1.

We first examined the modern coral record. As NADA PC1 represents seasonally averaged moisture variability that has its highest correlation with winter (January–March) tropical Pacific SSTs, we herein used the January–March mean coral $\delta^{18}$O records to make them directly comparable. All series were processed using a 9-yr Lanczos high-pass filter[30] to highlight interannual variability. The highest correlation of the modern coral $\delta^{18}$O record with NADA PC1 over their common period 1891–1994 is −0.58, which is close to its correlation with the Niño3 index (−0.61) in the same period. The highest correlation was achieved at lag zero, confirming that there is no dating error for modern coral. This good match between modern coral and NADA PC1 corroborates the dating error correction method.

The seventeenth-century sequence was spliced from three relict coral $\delta^{18}$O records (SB3, SB13, and SB8), and was correlated at −0.06 with NADA PC1 before dating errors were corrected. Dating error for each of the three individual series was corrected by adjusting years within the range of U/Th-dating errors. A firm match with NADA PC1 was achieved for each series after dating adjustments, with correlations improved from 0.17 to −0.41 for SB3, −0.20 to −0.62 for SB13, and −0.27 to −0.51 for SB8. After correcting the dating errors and re-splicing the three sequences together, a good match between relict corals and NADA PC1 was achieved, with a correlation of −0.47 over their common period 1637–1705 (Supplementary Fig. S4). Similarly, dating errors were corrected for the fourteenth–fifteenth-century sequence, and its correlation with NADA PC1 improved from −0.14 to −0.35 over the common period 1339–1443.

The tenth- and twelfth-century sequences are both single relict coral $\delta^{18}$O records. The best match between the tenth-century sequence and NADA PC1 was achieved after adjusting the start year of the relict coral sequence from AD 933 to AD 931, and their correlation over the common period 931–954 improved from 0.08 to −0.43. The correlation of the twelfth-century sequence with NADA PC1 over their common period 1154–1215 is −0.19, and there is no significant improvement when its dating was adjusted within the range of its U/Th-dating error. Therefore, we did not make a dating error correction for this sequence.

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## Author contributions

J.L., S-P.X. and E.R.C. contributed to data analysis. J.L., S-P.X., E.R.C., G.H., and R.D. contributed to writing the paper. All authors discussed the results and commented on the manuscript.

## Additional information

The authors declare no competing financial interests. Supplementary information accompanies this paper on www.nature.com/natureclimatechange. Reprints and permissions information is available online at http://npg.nature.com/reprintsandpermissions. Correspondence and requests for materials should be addressed to J.L.

Article

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# Spatial and temporal variations of light rain events over China and the mid-high latitudes of the Northern Hemisphere

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Based on observed daily precipitation data, monthly gridded radiosonde upper air temperature and sea surface temperature data from the UK Met Office Hadley Centre, monthly surface air temperature from the Climatic Research Unit at the University of East Anglia and the NCEP/NCAR monthly reanalysis data, this study investigates the spatial and temporal variations of light rain events over China and the mid-high latitudes of the Northern Hemisphere during 1961–2010, and discusses the relationship between the change of light rain events and atmospheric stability, sea surface temperature and atmospheric circulation. The light rain events over East China display a decreasing trend of 3.0%/10 a in summer and winter half years. Over Northwest China, an increasing trend of 4.1%/10 a is found in winter half years, but there is no trend in summer half years. Using empirical orthogonal function (EOF) analysis, it is found that the first two principal components of light rain events over the mid-high latitudes of the Northern Hemisphere show long time scale variations in summer and winter half years. The first EOF modes (EOF1s) for summer and winter half years both depict a long-term increase in light rain events over North America and Southern Europe as well as Northwest China (except in summer half years), and a long-term decrease over most of the Eurasia (Central Europe, Eastern Europe, North Asia and East China). The second EOF mode (EOF2) for summer half year shows that light rain events increase over North America, Southern Europe and South China, but decrease over Eurasia north of 45°N from 1961 to early 1980s, while the trends reverse from late 1980s to 2010. The second EOF mode (EOF2) for winter half years indicates that light rain events increase over North America and South and North China, but decrease over Eurasia north of 40°N from 1961 to early 1980s, while the trends reverse from late 1980s to 2009. Correlation analysis and linear regression analysis suggest that EOF1s may be related to the change in atmospheric static stability associated with global warming, and EOF2s are possibly linked to the AMO.

**light rain, the mid-high latitudes of the Northern Hemisphere, EOF, atmospheric stability, global warming**

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In recent decades, the global precipitation characteristic has changed significantly, following the continuous rise in the global temperature. In general, there have been increases in land precipitation at high latitudes, and deceases at the tropics and subtropics [1,2]. Precipitation in China as a whole has increased slightly since 1960s, with strong regional and seasonal differences [3–6]. Regionally, precipitation increased in Northwest China, but decreased in North China; seasonally, precipitation increased in winter and summer, but decreased in spring and autumn; precipitation extreme events increased in Northwest China and South China, but decreased in North China. Previous studies [7–12] mainly focused on the precipitation amount and extreme precipitation events, with little attention to the precipitation frequency, especially the light rain frequency. Trenberth [13] indicated that the characteristics of precipitation are just as vital as the total amount, and they are more apt to change as climate changes. Additionally, light rain events account for

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a large part of the precipitation events, while the proportion of extreme precipitation events is quite small [14,15]. Moreover, light rain soaks into the soil and benefit plants, while the heavy rain may cause local flooding and runoff, leaving soil much drier at the end [1,13]. Previous studies primarily investigated the long-term change of light rain in China. The results of Qian et al. [16] and Fu et al. [17] showed that the summer and annual light rain events (< 1 mm/d) in East China decreased, and the change might be associated with global warming. However, the study of Qian et al. [18] suggested that the light rain decrease in East China be a result of air pollution. Furthermore, Liu et al. [19] found that precipitation events of 0.1–0.3 mm/d decreased in whole China, and indicated that the reduction might be related with the decrease in solar radiation and cloud.

The changes of light rain events over large spatial region have not been well documented, although the decrease in light rain events in East China has been found. Therefore, this study is expected to analyze the spatial and temporal variations of light rain events over the Northern Hemisphere. Although Qian et al. [20] has investigated the long-term change of light rain over North America, Europe and Asia using Global Summary of Day (GSOD) dataset, the results have some uncertainties due to errors and a number of missing values in the dataset that consists of real-time data from Global Telecommunication System (GTS) [21,22]. This study investigates the spatial and temporal variations of light rain events over China and the mid-high latitudes of the Northern Hemisphere based on daily rain gauge precipitation data with quality control, and then discusses the probable causes of these changes.

## 1 Data and methods

### 1.1 Data

Data for this study consist of daily rain gauge precipitation dataset, monthly gridded radiosonde upper air temperature dataset [23] and sea surface temperature dataset [24] from the UK Met Office Hadley Centre (Hadley), monthly surface air temperature dataset [25] from the Climatic Research Unit at the University of East Anglia (CRU) and the NCEP/NCAR monthly reanalysis dataset [26].

Daily rain gauge precipitation data includes Global Historical Climatology Network-Daily (GHCN-Daily) station precipitation dataset [21] from the National Climate Data Center (NCDC), European station precipitation dataset [27] from European Climate Assessment and Dataset (ECA&D) project and Chinese station precipitation dataset from China Meteorological Administration (CMA). We replace the precipitation data over China and Europe in GHCN-Daily dataset with those from CMA and ECA&D project, since the GHCN-Daily dataset contain fewer stations in China especially in western China and more missing values in China and Europe especially for some stations in China after late 1990s and for some European stations after 2000.

Monthly gridded radiosonde upper air temperature dataset has a horizontal resolution of 10°×5°, 9 levels from 850 to 30 hPa, which is available starting in January 1958. Monthly sea surface temperature dataset has a horizontal resolution of 1°×1° and is available from January 1870 forward. Monthly surface air temperature has a horizontal resolution of 5°×5°, available from January 1850 forward. The NCEP/NCAR monthly reanalysis dataset includes sea level pressure and air temperature on pressure levels (17 pressure levels from 1000 to 10 hPa), with a horizontal resolution of 2.5°×2.5° and covering the period of January 1948 to present.

### 1.2 Data quality control

In order to ensure the reliability of the results, the daily precipitation dataset is quality controlled. NCDC has applied a set of quality assurance procedures to GHCN-Daily dataset to detect duplicate data, climatological outliers and various inconsistencies. A detail description was documented by Durre et al. [28]. We remove the data errors failing in quality checks and retain the valid data. ECA&D project has also checked the European station daily precipitation dataset and detected data errors including climatological outliers and negative values. We only use the data passing quality tests. Since the daily precipitation dataset from CMA hasn't been quality controlled by CMA, we detect the climatological outliers, negative values and world record exceedances referring to the quality checks applied to GHCN-Daily [28], and retain the valid data.

In this study, the year with no missing value is classified as usable. A station with no less than 40 usable years during 1961–2010 is accepted in this analysis. As a result, we select 3033 stations over the Northern Hemisphere, with 1135 stations from GHCN-Daily, 1276 stations from ECA&D and 622 stations from CMA. The locations of these stations are shown in Figure 1. Since most of the stations locate over the mid-high latitudes of the Northern Hemisphere with a small number of exception at low latitudes (0°–30°N), the stations at low latitudes have not been excluded in this study.

### 1.3 Methods

CMA defines light rain as that with daily precipitation less than 10 mm. The 10 mm/d was also selected as the threshold for light rain in Qian et al.'s [18] and Qian et al.'s [20] investigations of the changes of light rain events over East China and North America, Europe and Asia. According to this definition, almost all the stations' light rain events in climatological mean account for more than 50% of precipitation events over the mid-high latitudes of the North Hemisphere, with 50%–80% at mid latitudes and more than 90% at high latitudes and Northwest China (figure not shown). Therefore, a fixed threshold to define light rain may not be

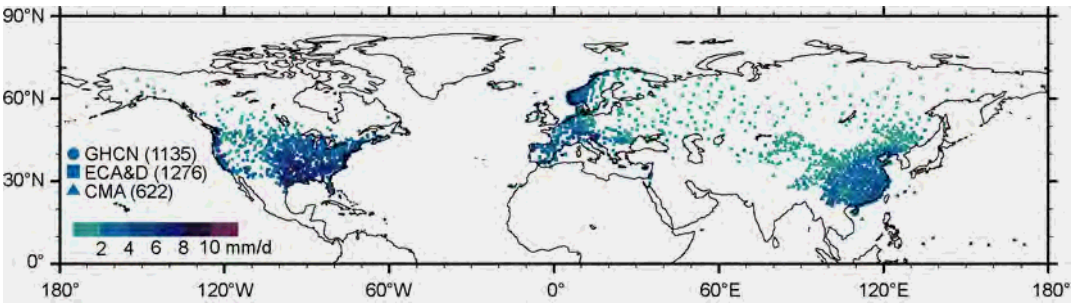



**Figure 1** The locations of 3033 precipitation stations and the thresholds for light rain events (the 50th percentile, mm/d). Circles, squares and triangles denote stations from GHCN-Daily, ECA&D, CMA, respectively.

appropriate. Considering regional differences in precipitation, we use percentile threshold to define light rain. The 50th percentile, which is the smallest percentage (47% actually) of precipitation days of less than 10 mm/d accounting for all the precipitation days in all stations, is chosen to be the threshold for light rain. Here, we want to point out that similar results are obtained when the analysis is based on the 30th or 40th percentile. For a station, if the daily precipitation is less than the threshold, then the precipitation event on that day is regarded as a light rain event. We refer to the method of Bonsal et al. [29] and Zhai et al. [30] to calculate the 50th percentile value. Daily precipitation during 1961–1990 is first ranked in ascending order $x_1$, $x_2$,$\cdots$, $x_m$, $\cdots$, $x_n$. The probability $P$ that a random value is less than or equal to the rank of that value $x_m$ is estimated by the equation of $p=(m-0.31)/(n+0.38)$, in which $m$ is the sequence number of $x_m$. The 50th percentile value is the $x_m$ when $m$ makes $p$ equal 50%. The thresholds for light rain event of every station are shown in Figure 1, ranging from 6–12 mm/d in Southeast United States, 2–6 mm/d in Southeast China and less than 2 mm/d over most of the mid-high latitudes of the Northern Hemisphere.

Considering the seasonal differences of climate background, this study divides a year into summer half year and winter half year, and analyzes the changes of light rain events in summer and winter half years respectively. Same as the study of Zhai et al. [6], summer half year is from April to September and winter half years is from October to March in the next year.

Linear regression is used to analyze the long-term trends in light rain events over China. Empirical orthogonal function (EOF) analysis is employed to obtain the first two dominant EOF modes and principal components of the light rain events over the mid-high latitudes of the Northern Hemisphere. Like the method of Wu et al. [31], the EOF modes and principal component, are multiplied by the square root of the eigenvalues and the reciprocal of the square root of eigenvalues, respectively. Linear regression and correlation analysis are applied to the investigations of the relationship between the change of light rain events and atmospheric stability, sea surface temperature and atmospheric circulation. The Student's *t*-test is used to examine the significance of the linear regression and correlation.

## 2 Results and discussion

### 2.1 Long-term trends in light rain events over China

Previous studies [16–18] mainly focused on the decrease in light rain events over East China, while the change in the rest of China was less studied. Moreover, fixed threshold was often used to define light rain. In this analysis, the changes of light rain events in summer and winter half years based on percentile thresholds are investigated over whole China.

Figure 2(a) shows the long-term trends in light rain events of 622 stations over China for summer half years during 1961–2010. Decreasing trends in light rain events are obvious in East China (east of 100°E). Although more stations in Northwest China (west of 100°E, north of 30°N) have upward trends, most of the trends are not statistically significant. The regionally averaged light rain events in East China has a decreasing trend of 0.9 d/10a or equivalently 3.0%/10 a (Figure 2(b)), which exceeds the 99% confidence level; while there is no significant tendency in regionally averaged light rain events over Northwest China (Figure 2(c)).

Figure 2(d) is the same as Figure 2(a) but for the winter half years during 1961–2009. The light rain events for winter half years in East China primarily exhibit decreasing trends, although the trends for some stations over northern China are not statistically significant and a small number of stations in Northeast China show increasing trends. However, the light rain events for most of the stations in Northwest China have increasing trends. The trend in regionally averaged light rain events in East China is −0.8 d/10 a corresponding to −3.0%/10 a (Figure 2(e)), and that in Northwest China is 0.6 d/10 a or 4.1%/10 a, which both exceed the 99% confidence level (Figure 2(f)).

### 2.2 Spatial and temporal variations of light rain events over the mid-high latitudes of the Northern Hemisphere

The EOF analysis is applied to the light rain days of 3303

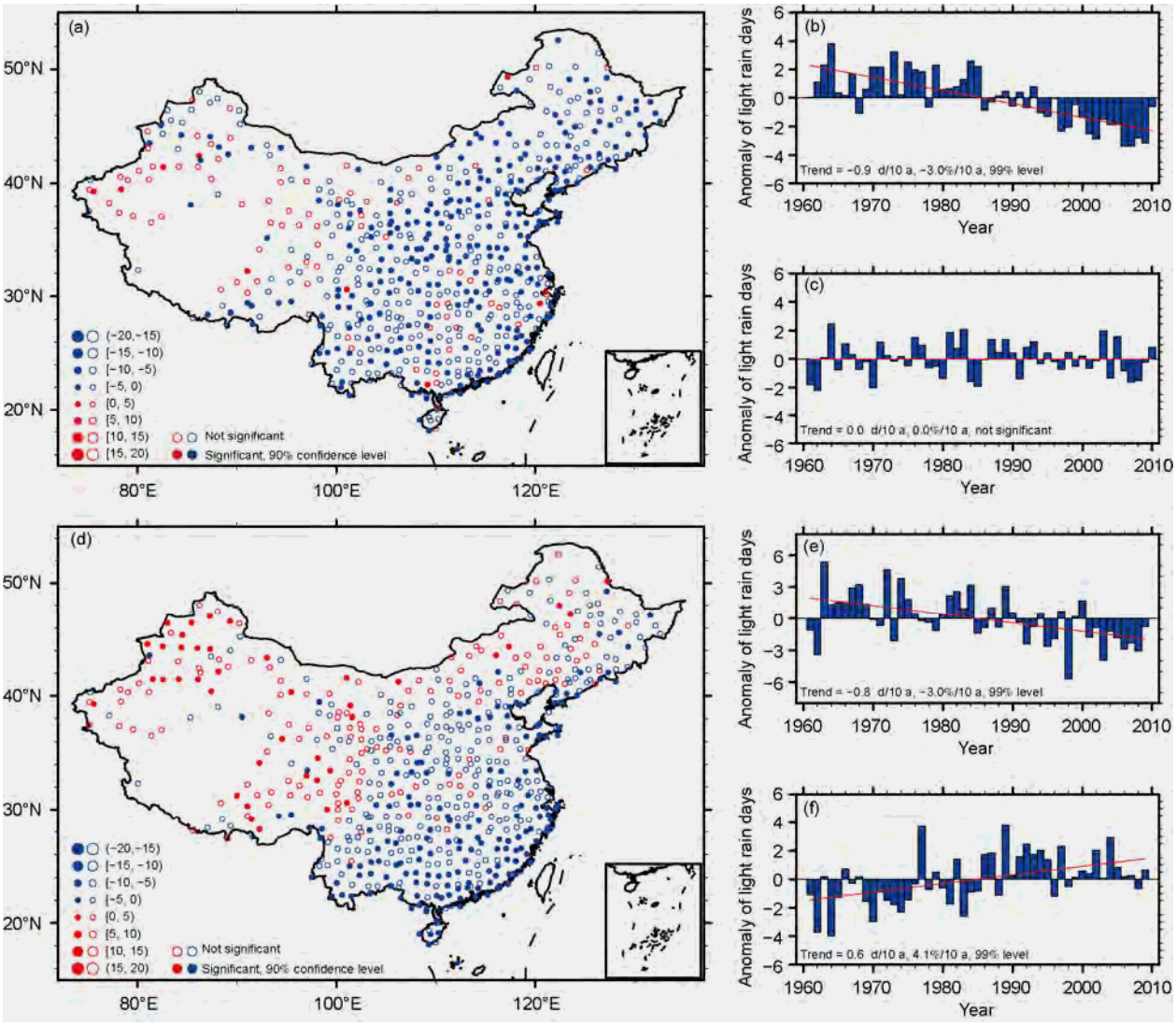



**Figure 2** Trends (d/10 a) in light rain days of 622 stations over China (a) and light rain day anomalies (d) averaged over East China (east of 100°E) (b) and Northwest China (c) in summer half years during 1961–2010. (d), (e) and (f) are the same as (a), (b) and (c), respectively, but for winter half years.

stations over the mid-high latitudes of the Northern Hemisphere in summer and winter half years during 1961–2009 respectively. The first two EOF modes and the corresponding principal components for summer and winter half years, which explain about 10.2% and 5.2% of the total variances for summer half years and 10.3% and 6.5% for winter half years, are shown in Figure 3.

The first two principal components (PCs) for summer half years indicate slow variations of light rain events (Figure 3(c)). PC1 represents the long-term trends during 1961–2010 and, PC2 depicts the two opposite trends during 1961 to early 1980s and during late 1980s to 2010. The first EOF mode (EOF1, Figure 3(a)) for summer half years is characterized by a positive anomaly covering North America, Southern Europe and a negative anomaly covering most of the Eurasia including Central and Eastern Europe, North Asia (east of the Urals, about east of 60°E) and East China. As mentioned above, the corresponding principal component (PC1) represents the long-term trends in recent 50 years. Therefore, the EOF1 and PC1 indicate that the light rain events in summer half years increase over North America and Southern Europe during 1961–2010, but decrease over most of the Eurasia (Central and Eastern Europe, North Asia and East China), especially in East China. The second EOF mode (EOF2, Figure 3(b)) for summer half years is characterized by a negative anomaly in North America, Southern Europe and South China and a positive anomaly over Eurasia north of 45°N. The second principal component (PC2) depicts a decreasing trend from 1961 to early 1980s and an increasing trend from late 1980s to 2010. Hence, the EOF2 and PC2 reveal that the light rain events in summer half years increase in North America, Southern Europe and South China during 1961 to early 1980s, but decrease over Eurasia north of 45°N; the trends reverse in these areas during late 1980s to 2010.

The leading two principal components of light rain events

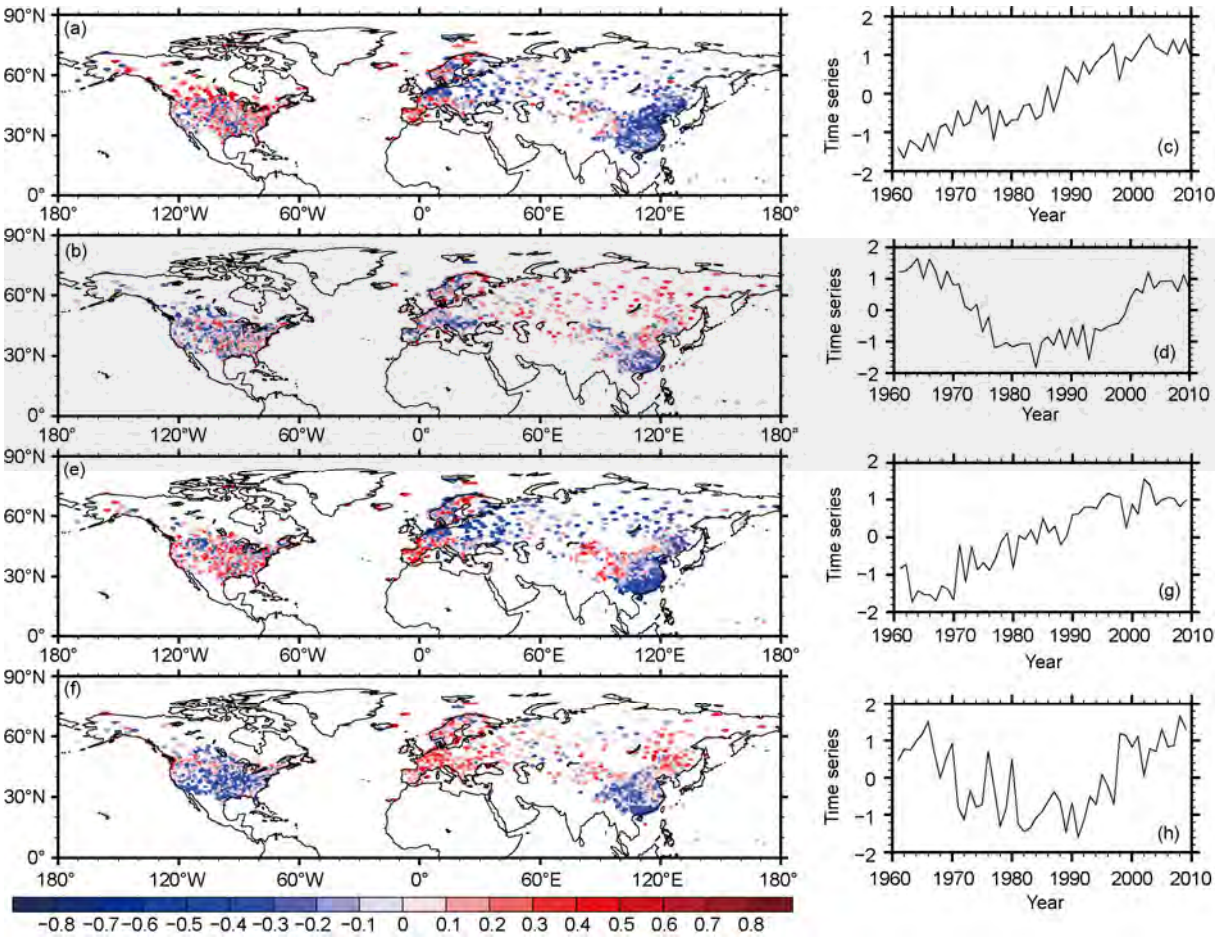



**Figure 3** The first EOF mode (a) and the second EOF mode (b) of light rain days over the mid-high latitudes of the North Hemisphere and the first principal components (c) and the second principal components (d) for summer half years during 1961–2010. (e), (f), (g) and (h) are the same as in (a), (b), (c) and (d), respectively, but for winter half years.

in winter half years are similar to those for summer half years respectively, with PC1 (Figure 3(g)) depicting long-term change and PC2 (Figure 3(h)) describing opposite trends during periods of 1961 to early 1980s and late 1980s to 2009. The first EOF mode (EOF1, Figure 3(e)) for winter half years is also similar to that for summer half years except for Northwest China where the long-term change of light rain events is significant. The EOF1 for winter half years is characterized by a positive anomaly covering North America, Southern Europe and Northwest China and a negative anomaly covering most of the Eurasia (Central and Eastern Europe, North Asia and East China). This indicates that the light rain events in winter half years increase over North America, Southern Europe and Northwest China during 1961–2010, but decrease over most of Eurasia (Central and Eastern Europe, North Asia and East China). The second EOF mode (EOF2, Figure 3(f)) for winter half years is analogous to that for summer half years except for Southern Europe, where the change for winter half years is contrary to summer half years. The EOF2 for winter half years shows a negative anomaly in North America, South and North China and a positive anomaly over Eurasia north of 40°N. This implies that the light rain events in winter half years increase in North America, South and North China during 1961 to early 1980s, and decrease over Eurasia north of 40°N; the trends reverse in these areas during late 1980s to 2009.

### 2.3 Discussion

Precipitation events are generally accompanying upward motions. The vertical motion is usually associated with the atmospheric stratification, in other words, it is associated with the vertical distribution of atmospheric temperature and humidity. Unstable atmospheric stratification favors upward motion and heavy precipitation, while stable atmospheric stratification suppresses the upward motion and is unfavorable for intense precipitation [32]. Thus the change of atmospheric static stability may influence the change of light rain events. While the atmospheric stability strengthens, the upward motion abates so that the precipitation intensity decreases and light rain events increase. When the atmospheric stability weakens, the upward motion strengthens so that the precipitation intensity increases and

light rain events decrease. As mentioned above, the first principal component of light rain events in both summer and winter half years represent the long-term trend during 1961–2010. This section mainly investigates the relationship between the trend of light rain events and the atmospheric stability over the mid-high latitudes of the Northern Hemisphere.

Figure 4(a) and (b) shows the trend of lower troposphere temperature difference between lower and upper level in summer half years during 1961–2010, based on reanalysis air temperature from the NCEP/NCAR and upper radiosonde temperature from Hadley Centre, surface air temperature from CRU, respectively. The result from NCEP/NCAR reanalysis air temperature (Figure 4(a)) indicates that the temperature difference between 1000 and 700 hPa decrease over the mid latitudes of North America in recent 50 years, where the temperature difference between surface air temperature from CRU and radiosonde temperature at 700 hPa from Hadley Center (Figure 4(b)) also shows a decreasing trend. Thus, both datasets imply that the atmospheric stability strengthens over the mid latitudes of North America. As mentioned above, the light rain events for summer half years increase over North America. The air temperature differences between lower level and upper level decrease and the atmospheric stability weakens over Central and Eastern Europe, North Asia and East China (Figure 4(a), (b)), where the light rain events decrease. The air temperature differences in winter half years from reanalysis data (Figure 4(d)) and from CRU data and Hadley Centre radiosonde data (Figure 4(e)) have decreasing trends over the mid latitudes of North America and Southern Europe (not shown in Figure 4(d) for statistical significance). This indicates that the atmospheric stability strengthens. The air temperature differences in winter half years have increasing trends over Central and Eastern Europe, North Asia and East China (except for South China in reanalysis data) based on CRU and Hadley Centre data (Figure 4(e)) and the NCEP/NCAR reanalysis data (Figure 4(d)). This corresponds to the light rain events decrease over Central and Eastern Europe, North Asia and East China. The reanalysis data (Figure 4(d)) show that the temperature differences between lower and upper level in winter half years decrease and the atmospheric stability strengthens over South and Northwest China, while the observed data (Figure 4(e)) indicate that the atmospheric stability weakens over South and Northwest China. The contrary results from different data make it difficult to verify the association between the change of light rain events and atmospheric stability over South and Northwest China. From the above analysis, in most of the areas light rain events increase when atmospheric stability strengthens and light rain events decrease when atmospheric stability weakens. However, there are still some exceptions. For example, the air temperature difference between surface and 700 hPa in summer half years based on observed data increase and atmospheric stability weakens over Southern Europe (Figure 4(b)), but the light

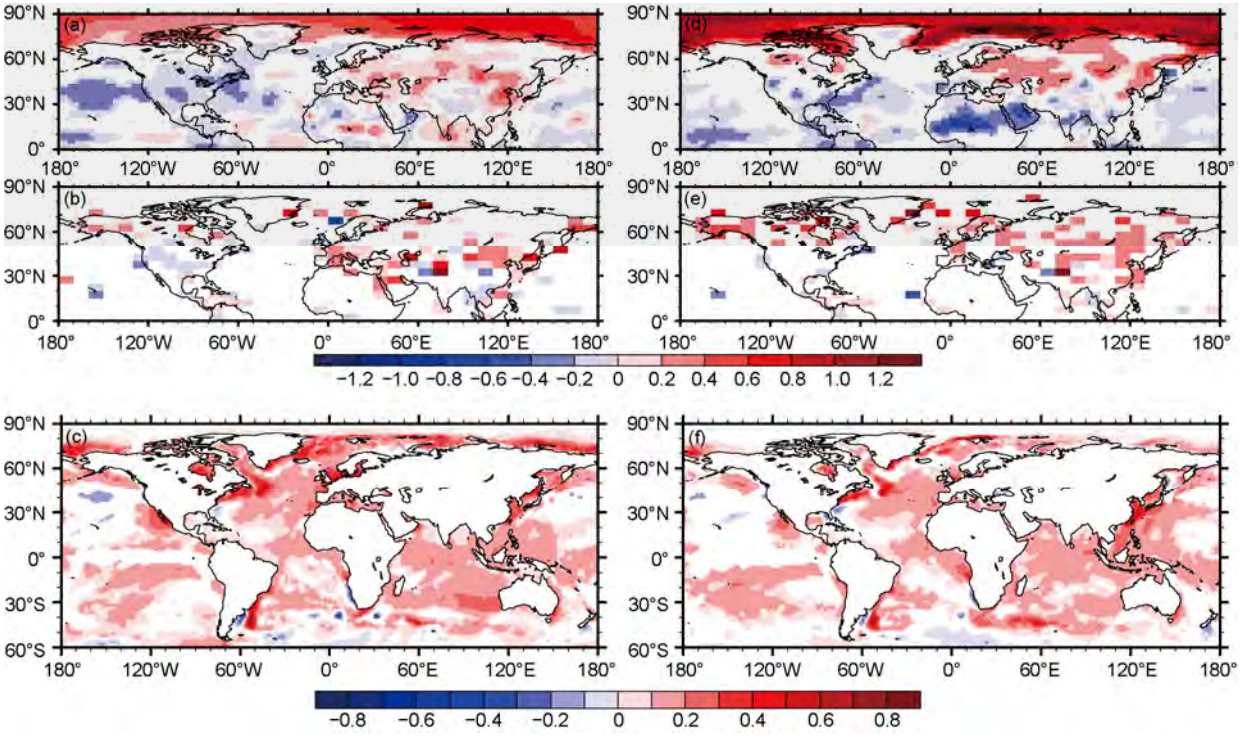



**Figure 4** Trends (°C/10 a) in air temperature difference between 1000 and 700 hPa based on NCEP/NCAR Reanalysis air temperature (a) and between surface and 700 hPa based on surface air temperature from CRU and radiosonde upper air temperature from Hadley Centre (b), and trends (K/10 a) in global sea surface temperature (c) in summer half years during 1961–2010. (d), (e) and (f) are the same as in (a), (b) and (c), respectively, but for winter half years. Shaded areas denote 90% confidence level.

rain events increase. So the change may be influenced by other factors. This study primarily discusses the probable causes for the change of light rain events on large spatial scale, and the local factors are not our concern in this study. The cause for the change of precipitation pattern on local scale may be complicated. We will investigate the local cause in the near future.

The trends in sea surface temperature for summer and winter half years are shown in Figure 4(c) and Figure 4(f), respectively. A number of studies [2,33–35] have pointed out that the Earth surface (sea and land surface) and troposphere temperatures continuously increase. Following the increase in sea surface temperature, the evaporation increases and more water vapor would be in the atmosphere. The latent heat from more water vapor condensation makes the upper air become warmer. Meanwhile, the Earth surface warming makes the surface air temperature increase. However, the magnitudes of warming of lower and upper air are different in different area. So the atmospheric stability changes, which induce the change in light rain events.

In addition, the present analysis indicates an increase in light rain events for summer and winter half years over North America and Southern Europe, which is different from the result of Qian et al. [20] who showed that the light rain events decrease over North America and Europe (including Southern Europe). Their results may be associated with too high threshold for light rain event and too much missing values (less than 10% of annual days, which is 36 d/a) in the used data.

A number of studies [36–42] have revealed the connection between the precipitation or extreme precipitation events and sea surface temperature and atmospheric circulation. In order to investigate the relationship between the second EOF mode of light rain events and sea surface temperature and atmospheric circulation, we calculate respectively the correlation coefficient between the PC2 of the light rain events in summer and winter half years and global sea surface temperature (SST) and sea level pressure (SLP), which are shown in Figure 5. It can be seen from Figure 5(a) and (c) that high correlation between PC2 and SST are mainly located in North Atlantic and North Pacific in summer and winter half years. North Atlantic sea surface temperature has multidecadal variation, which is known as Atlantic Multidecadal Oscillation (AMO). The AMO can impact the climate of North America, Europe and Asia [43–46]. We calculate the correlation coefficient between AMO index [47] and PC2 for summer and winter half years. The coefficients are 0.56 and 0.76, respectively, exceeding the 99% confidence level (correlation coefficient of 0.36).

The observation and simulation results of previous studies [45,46,48] showed that during the AMO warm phase, the summer rainfall and river flows decrease in North America, while summer precipitation increase in Europe, which implies that there are negative correlation between the AMO and summer precipitation in North America and positive correlation in Europe. The AMO was changing from warm phase to cold phase during 1961 to early 1980s and from cold phase to warm phase after late 1980s (figure not shown), which is similar to the change of PC2. As mentioned above, the second EOF mode of light rain events in summer half years indicates that light rain events increase over North America and decrease over Europe (except for Southern Europe) during 1961 to early 1980s, while the trends reverse after late 1980s. Therefore, there are negative

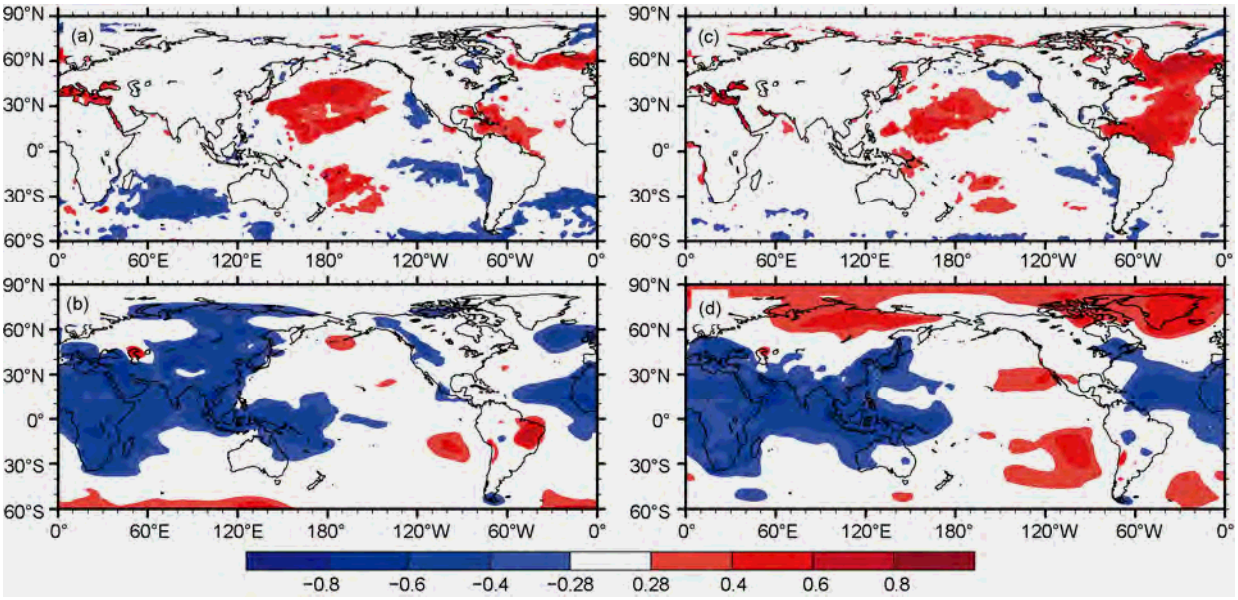



**Figure 5** Correlation coefficient between the second principal component of the light rain days and sea surface temperature (a) and sea level pressure (b) in summer half years during 1961–2010. (c) and (d) are the same as in (a) and (b), respectively, but for winter half years. Shaded areas denote 95% confidence level.

correlation between the AMO and the second EOF mode of light rain events in summer half years over North America and positive correlation over most of Europe. Sutton et al. [45] found that there are two low-pressure anomaly centers over North Atlantic during the AMO warm phase, one located to the southwest of United States, the other being to the west of Europe, and indicated that the two anomaly centers were associated with the precipitation change over North America and Europe. It can be seen from the pattern of the correlation between PC2 and sea level pressure (Figure 5(b)) that there are two significant correlative regions over North Atlantic, with the northern one corresponding to the anomaly center to the west of Europe, and the southern one located in the North Atlantic to the southwest of United States, to the southeast of the anomaly center. Thus, the AMO is probably linked to the second EOF mode of light rain events for summer half years over North America and Europe. The AMO influences the change of precipitation and precipitation days, and the change of light rain events probably keeps in line with the change of precipitation and precipitation days. Therefore, the second EOF mode of light rain events displays opposite trends in two periods. However, whether the change of light rain events is accordance with precipitation and precipitation days or not should be investigated further. Using observed precipitation data, Li et al. [44] and Wang et al. [43] found that during the AMO warm phase, the autumn and winter precipitation decrease in South China. The result indicates that there is negative correlation between the AMO and autumn and winter precipitation in South China, which is in accordance with the fact that South China is covering negative anomaly in EOF2 for winter half years. However, the mechanism how AMO influences the East Asian climate is not clear, and should be studied onward [49]. Although no research revealed the connection between the AMO and autumn and winter precipitation over North America and Europe, Knight et al.'s model results [48] showed that the AMO was directly related to autumn and winter precipitation over Europe, in agreement with the fact that Europe is covering positive anomaly in EOF2 for winter half years. Their results also indicated that the AMO was negatively correlated with the autumn and winter sea level pressure, which is similar to the pattern of correlation between PC2 and sea surface pressure in winter half years (Figure 5(d)). Thus, the AMO is possibly associated with the second mode of light rain events in winter half years. It should be noted that the AMO is a multidecadal oscillation with a cycle of 65–80 years [49], but the temporal period in this study is only 50 years, which is less than a cycle of the AMO. Therefore, longer precipitation data is needed to verify the association between the AMO and the second EOF mode of light rain events.

The North Pacific sea surface temperature decreased during 1961 to early 1980s, and increased after late 1980s (figure not shown). Thus PC2 of light rain events in summer and winter half years is positively correlated with the North Pacific sea surface temperature. Nevertheless, we have not found the probable connection between the North Pacific sea surface temperature and the second EOF mode of light rain events. Aleutian low weakens during North Pacific warm phase and strengthens during cold phase [50]. Nevertheless, there is no significant correlation between PC2 for summer and winter half years and Aleutian low (Figure 5(b), (d)). In addition, the precipitation in coastal Southeast China increases during warm phase and decreases during cold phase [51,52]. The positive correlation disagrees with the negative anomaly in coastal Southeast China in EOF2 (Figure 3(b),(f)). Thus, it could not be determined that the change of light rain events is in accordance with the change of precipitation and precipitation days.

## 3 Summary

This study investigates the long-term trend in light rain events over China in summer and winter half years during 1961–2010 based on daily precipitation data of 622 stations. The results show that the light rain events over East China in summer and winter half years decrease significantly by 0.9 d/10 a and 0.8 d/10 a corresponding to 3%/10 a respectively, while those in Northwest China increase by 0.6 d/10 a corresponding to 4.1%/10 a in winter half years, but display no trend in summer half years.

In order to study the change of light rain events on large spatial scale, this investigation uses EOF analysis to obtain the first two EOF modes and corresponding principal components of light rain events in summer and winter half years, based on station daily precipitation data over the mid-high latitudes of the Northern Hemisphere. The first two principal components (PCs) for summer and winter half years are characterized by long time scale changes, with PC1 for summer and winter half years both representing long-term trend during 1961 to 2010 and PC2 for summer and winter half years both standing for opposite trends during periods of 1961 to early 1980s and late 1980s to 2010. The first EOF modes (EOF1) for summer and winter half years both primarily depict that the light rain events increase over North America and Southern Europe and decrease over most of Eurasia (including Central and Eastern Europe, North Asia and East China) during 1961–2010. The EOF1 for winter half years also shows that the light rain events increase in Northwest China. The second EOF mode (EOF2) for summer half years reveals that the light rain events increase over North America, Southern Europe and South China during 1961 to early 1980s, and decrease over Eurasia north of 45°N, while the trends reverse after late 1980s. The second EOF mode (EOF2) for winter half years indicates that the light rain events increase over North America, South and North China during 1961 to early 1980s, and decrease over Eurasia north of 40°N, while the trends reverse after late 1980s.

This study discusses the probable cause for the change of light rain events. The relationship between the pattern of the trend in atmospheric stability and the trend in light rain events depicted in EOF1 indicate that in most areas the light rain events increase when atmospheric stability strengthens, and decrease when atmospheric stability weakens. The long-term change of atmospheric stability is a probable cause for the long-term change of light rain events. The change of atmospheric stability is possibly related to different vertical warming of troposphere associated with global warming. The second EOF modes may be linked to the AMO. Nevertheless, the influencing mechanism is not clear, and needs to be investigated further.

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# Strengthening of Tropical Indian Ocean Teleconnection to the Northwest Pacific since the Mid-1970s: An Atmospheric GCM Study*

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ABSTRACT

The correlation of northwest (NW) Pacific climate anomalies during summer with El Niño–Southern Oscillation (ENSO) in the preceding winter strengthens in the mid-1970s and remains high. This study investigates the hypothesis that the tropical Indian Ocean (TIO) response to ENSO is key to this interdecadal change, using a 21-member ensemble simulation with the Community Atmosphere Model, version 3 (CAM3) forced by the observed history of sea surface temperature (SST) for 1950–2000. In the model hindcast, the TIO influence on the summer NW Pacific strengthens in the mid-1970s, and the strengthened TIO teleconnection coincides with an intensification of summer SST variability over the TIO. This result is corroborated by the fact the model's skills in simulating NW Pacific climate anomalies during summer increase after the 1970s shift.

During late spring to early summer, El Niño–induced TIO warming decays rapidly for the epoch prior to the 1970s shift but grows and persists through summer for the epoch occurring after it. This difference in the evolution of the TIO warming determines the strength of the TIO teleconnection to the NW Pacific in the subsequent summer. An antisymmetric wind pattern develops in spring across the equator over the TIO, and the associated northeasterly anomalies aid the summer warming over the north Indian Ocean by opposing the prevailing southwest monsoon. In the model, this antisymmetric spring wind pattern is well developed after but absent before the 1970s shift.

## 1. Introduction

The northwest Pacific (NWP) summer monsoon is an important component of the Northern Hemisphere summer monsoon (Murakami and Matsumoto 1994; Ueda and Yasunari 1996; Ueda et al. 2009; Wu and Wang 2000). Year-to-year variations of the NWP monsoon significantly affect the occurrence of the flooding–drought events over East Asia (Huang and Wu 1989; Chang et al. 2000) via the Pacific–Japan teleconnection pattern (Nitta 1987; Kosaka and Nakamura 2006; Arai and Kimoto 2008). Interannual variability in the NWP summer monsoon is correlated with El Niño–Southern Oscillation (ENSO) in the preceding winter (Wang et al. 2003). In the summer following El Niño, rainfall decreases over the subtropical NWP with an anomalous anticyclone near the surface. In East Asia, summer rainfall tends to be excessive.

ENSO-induced SST anomalies persist through summer over the tropical Indian Ocean (TIO; Du et al. 2009). They have been implicated as anchoring summer climate anomalies over the NWP and East Asia via the so-called capacitor effect (Yang et al. 2007; Li et al. 2008). Xie et al. (2009) show that atmospheric Kelvin waves are

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the key bridge between the TIO and subtropical NWP. El Niño–induced TIO swarming force a Matsuno (1966)–Gill (1980) pattern in tropospheric temperature. The warm tropospheric Kelvin wave propagates into the western Pacific, inducing Ekman divergence in the subtropical NWP to suppress convection and develop an anticyclonic circulation near the surface. Recent forecast experiments using a coupled general circulation model (GCM) confirm this TIO teleconnection to the NWP (Chowdary et al. 2010).

The correlation between the NWP monsoon and ENSO has apparently strengthened since the mid-1970s, possibly related to the enhanced ENSO variance (Wang et al. 2008). An observational study by Xie et al. (2010) suggests that this interdecadal change originates in changes in the TIO teleconnection to the NWP. The El Niño–induced TIO warming takes place during epochs both before and after the mid-1970s shift, but it persists into summer only after the mid-1970s. Accordingly, the tropospheric Kelvin wave it forces strengthens substantially in the postshift epoch.

The earlier-mentioned observational studies suggest a strengthening of the TIO teleconnection to the NWP, which seems to explain the apparent increase in correlation of NWP summer monsoon variability with ENSO two seasons in advance. Given that TIO SST anomalies are an important source of summer climate predictability over the NWP and East Asia (Chowdary et al. 2010), it is important to investigate the cause of the interdecadal change in the TIO teleconnection. While increased ENSO activity (Wang et al. 2008) and enhanced TIO SST variability (Xie et al. 2010) are a plausible cause, alternative explanations cannot be ruled out from observations alone. For example, improvements in observations and their assimilation into reanalysis products reduce sampling errors and may lead to an apparent increase in correlation, or chaotic variability in the atmosphere can also cause apparent variations in correlation.

The present study examines the cause of the apparent strengthening of the TIO teleconnection to the summer NWP based on a 21-member ensemble simulation using an atmospheric general circulation model (AGCM) forced by observed SST for 1950–2000. The AGCM ensemble simulation reproduces the interdecadal change similar to observations. We then proceed to examine interdecadal change in TIO SST response to ENSO and its subsequent influence on the NWP. The enhanced persistence of TIO SST anomalies since the mid-1970s is related to a change in wind pattern in the preceding season.

The rest of the paper is organized as follow. Section 2 describes the data and numerical simulation. Section 3 reviews observational evidence for the interdecadal change in the TIO teleconnection. Sections 4 and 5 diagnose the model output, examining the interdecadal change during boreal summer and spring, respectively. Section 6 is a summary.

## 2. Data and model

We use the Met Office Hadley Centre Sea Ice and SST (HadISST) dataset (Rayner et al. 2003) and the National Centers for Environmental Prediction (NCEP) atmospheric reanalysis (Kalnay et al. 1996), originally on 1° and 2.5° grids, respectively. Our study focuses on interannual variability. Except in the wavelet analysis, a linear trend has been subtracted from raw anomalies to remove decadal and longer variations.

SST anomalies averaged over the region of 10°S–20°N, 40°–100°E are used as an index for TIO SST variations, and the meridional difference of 850-hPa zonal wind anomalies between a southern region (5°–15°N, 100°–130°E) and a northern region (20°–30°N, 110°–140°E) is used as an index for the NWP summer anticyclone, following the NWP summer monsoon index by Wang et al. (2001), but with the sign reversed. Both indices are defined for boreal summer [June–August (JJA)]. Hereafter, seasons refer to those for the Northern Hemisphere.

We use the Community Atmosphere Model, version 3 (CAM3) at T42 resolution (equivalent to 2.8° latitude × 2.8° longitude) in the horizontal and 23 sigma levels in the vertical. Details of the models are provided in Collins et al. (2006). A 21-member ensemble of CAM3 simulations are performed with HadISST observations as surface boundary conditions for a 51-yr period of 1950–2000. Each member simulation differs only in the initial conditions.

## 3. Observed interdecadal change in TIO teleconnection

This section presents results from an observational analysis to be compared with the AGCM hindcast. Figure 1 shows the correlation in 15-yr sliding windows between JJA TIO SST and NWP anticyclone indices (solid line). The correlation is weak (−0.4 to 0.1) and insignificant for the first half of the record and rises dramatically in the mid-1970s. It peaks above 0.8 in 1980 and somewhat decreases afterward but stays above the 95% significance level for the past three decades. The strengthening in the mid-1970s of this TIO teleconnection to the NWP is the focus of this study. A weakening is also seen in the WNP anticyclone–TIO SST relationship in the mid-1990s, which may be related to the decadal change in the East Asian summer monsoon as studied by Kwon et al. (2005).

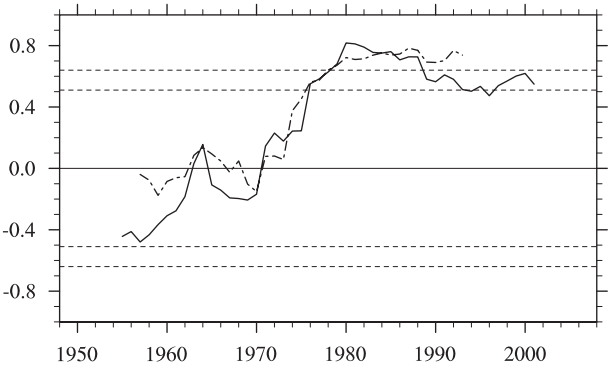



FIG. 1. The 15-yr sliding correlation between the NWP anticyclone and JJA TIO SST indices of observation (solid line), and for the NWP anticyclone index between observations and the ensemble-mean simulation (dot–dash line). Horizontal black dash lines denote 95% and 99% confidence levels.

To document the interdecadal change, we partition the data into preshift (PRE) and postshift (POST) periods of 1953–76 and 1977–2000, respectively. Figure 2 compares the correlations of SST, 850-hPa wind velocity, and tropospheric (850–200-hPa mean) temperature with the NWP anticyclone index between the PRE and POST epochs. During the POST epoch, positive SST correlation is found over the north TIO through the South China Sea. In response to the TIO warming, troposphere temperature displays a Matsuno–Gill pattern with the maximum correlation over the eastern TIO. A Kelvin wave wedge penetrates the western Pacific along the equator, with a pronounced anticyclonic circulation on its northern flank. These results are consistent with the Kelvin wave–induced Ekman divergence mechanism of Xie et al. (2009) for the NWP anticyclone. During the PRE epoch, SST correlation with the NWP anticyclone is weak, except over small patches off Sumatra, in the South China Sea and NWP. SST correlation features an east–west contrast in the TIO, with negative values in the western TIO and westerly wind anomalies on the equator. Consistent with weak SST anomalies, there is no significant tropospheric temperature correlation over the TIO. The NWP anticyclone correlation with the concurrent summer Niño-3.4 SST index is low, $-0.41$ and $-0.11$ for the PRE and POST epochs, respectively. These results support that the TIO influence on the NWP anticyclone is weak prior to the mid-1970s and strengthens afterward.

Figure 3 show wavelet spectra of summer TIO SST and NWP anticyclone indices. In a frequency band of approximately four years, TIO SST variability experienced a prominent increase in the late 1970s (Fig. 3a). There is a concurrent increase in NWP anticyclone variability in the same frequency band (Fig. 3b). Moreover, both TIO SST and NWP anticyclone variability displays a concurrent shift in dominant frequency from 4 to 3 yr in the 1990s. All this suggests TIO SST and its teleconnection as the cause of intensified variability in the NWP anticyclone. This hypothesis is consistent with the earlier-mentioned correlation analysis and will be tested with the CAM hindcast.

## 4. Summer interdecadal change in model

We first evaluate model skills by performing the empirical orthogonal function (EOF) analysis for ensemble-mean variability in JJA 850-hPa wind velocity over the NWP. Figure 4 compares the leading EOF modes with observations. In observations, the first mode explains 28% of the total variance, featuring an anticyclone associated with positive sea level pressure (SLP) anomalies over the subtropical NWP (Fig. 4, upper left). The second mode explains 16% of the variance, featuring westerly anomalies along the equator with zonal SLP gradient. There is a weak subtropical anticyclone over East Asia and the NWP (Fig. 4, lower left). Correlation analysis with eastern Pacific SST indicates that the second EOF (EOF-2) represents anomalies during the summer when El Niño develops [JJA(0)] and EOF-1 represents the summer after El Niño decays [JJA(1)]. Here the numerals in parentheses denote the El Niño developing (0) and decay (1) years. The first principal component (PC-1) is highly correlated with the NWP anticyclone index at $r = 0.92$.

The model ensemble-mean simulation captures the salient features of observed EOFs with the following differences: 1) the model EOF-1 explains a much larger fraction of variance possibly because of weaker noise than in observations; 2) the anticyclone in the model EOF-1 is shifted by 5° northward; and 3) in the model EOF-2, the subtropical anticyclone is confined to East Asia and the off-equatorial low-pressure belt to the south is more pronounced.

Figure 5 shows the 15-yr sliding correlation for PCs between observations and the ensemble mean. While the PC-2 correlation remains nearly constant and above the 95% significance level throughout, the model's skills in PC-1 displays a pronounced interdecadal change in the late 1970s, low and insignificant before the shift but rising rapidly afterward to values as high as 0.7. The ensemble-mean AGCM simulation helps isolate the SST-forced variability (Rowell et al. 1995; Lu et al. 2006). The high correlation for the NWP EOF-1 mode between observations and the ensemble-mean simulation during the POST epoch (Fig. 5) indicates that the year-to-year variations in the NWP summer monsoon are forced by SST anomalies, while the low correlation suggests that the SST forcing is weak during the PRE

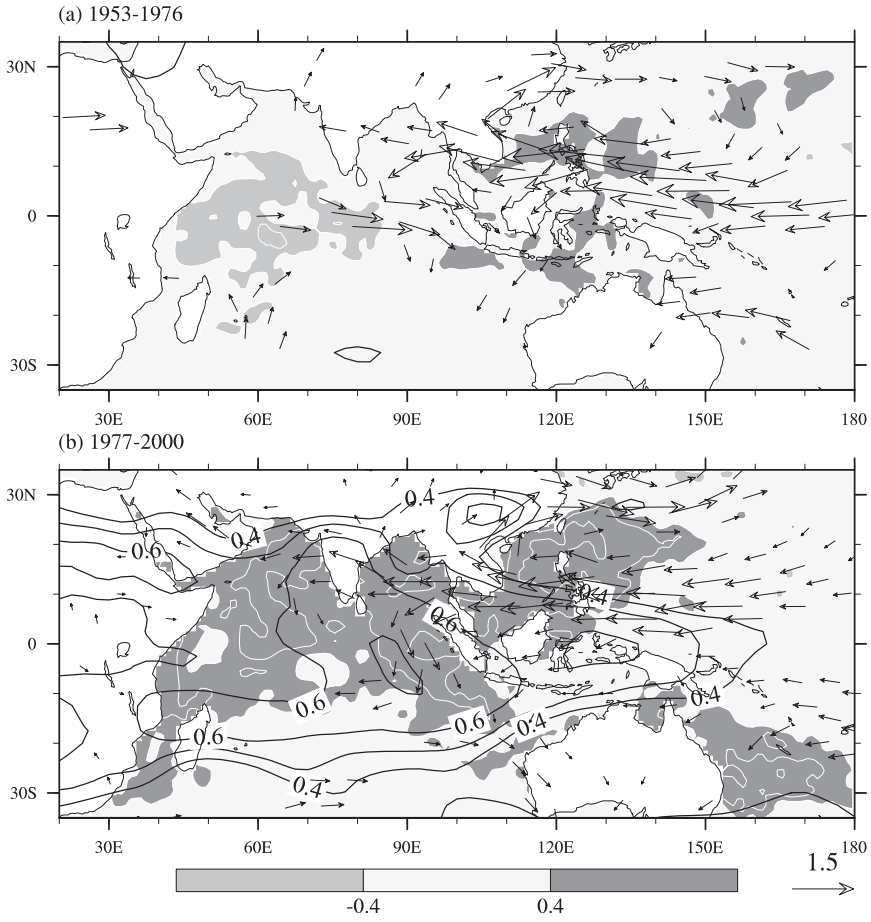



FIG. 2. Correlation with the NWP anticyclone index in observation during (a) 1953–76 and (b) 1977–2000: JJA SST (gray shade and white contours at intervals of 0.2) and tropospheric temperature (vertical average from 850 to 200 hPa; contours $r \geq 0.4$). Wind vectors (m $s^{-1}$, pass the 95% confidence level) show the regression of 850-hPa wind with the NWP anticyclone index.

epoch. Thus, it is possible that changes in SST patterns cause this interdecadal change.

Indeed, the model simulates the late 1970s' increase in interannual variability of the NWP anticyclone (Fig. 3c), even the shift of the dominant frequency from 4–5 to 3 yr. The concurrent increase in TIO SST variability suggests it as the cause of enhanced NWP anticyclone variability. NWP anticyclone variability may be due to internal variability or forced by SST anomalies (Lu et al. 2006). Ensemble AGCM simulations capture only SST-forced variability, which appears to dominate the NWP anticyclone after the mid-1970s. Figure 5 shows that the rise of the model's skills in simulating NWP climate anomalies coincides with the increase in observed correlation between TIO SST and NWP anticyclone variability (Fig. 2).

### *TIO teleconnection*

The rest of the section examines the TIO teleconnection to the NWP. Figure 6 compares the correlation between the TIO SST and NWP anticyclone in 21 simulations with CAM3. A total of 12 members out of 21 show the correlation exceeding the 95% confidence level during the POST epoch, but none passes the 95% confidence level during the PRE epoch. Moreover, the correlation coefficients are higher during the POST epoch than the PRE epoch in every member of the simulations, illustrating the strengthened TIO teleconnection after the mid-1970s.

Figure 7 compares summer correlations of ensemble-mean precipitation, tropospheric temperature, and 850-hPa wind with TIO SST between the PRE and POST epochs. Consistent with observations during the POST epoch, a TIO warming forces a Matsuno–Gill pattern in tropospheric temperature, with a warm Kelvin wave propagating into the western Pacific along the equator and Rossby wavelike off-equatorial maxima over the western TIO and Africa. Precipitation increases over the TIO except over the Bay of Bengal. A low-level anticyclone

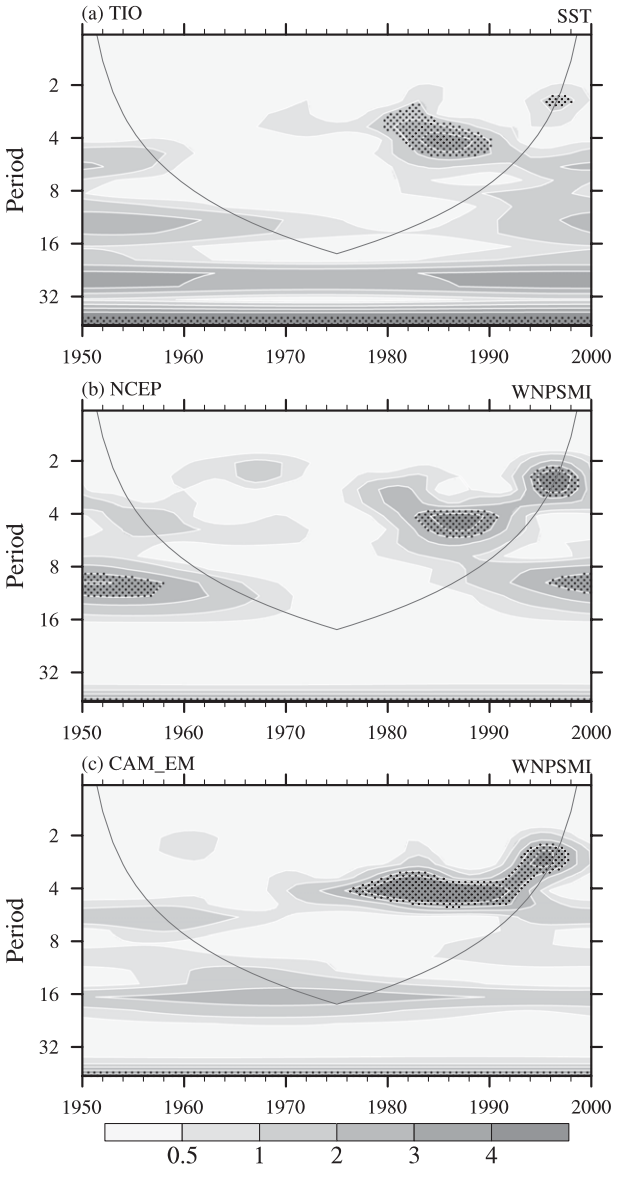



FIG. 3. Wavelet spectra of the standard deviation of the (a) JJA TIO SST index, (b) observations NWP anticyclone index, and (c) ensemble-mean NWP anticyclone index. All shaded regions denote 95% confidence levels.

appears on the northern flank of the tropospheric Kelvin wave from the Bay of Bengal to the subtropical WNP, where precipitation shows a tendency to decrease. During the PRE epoch, by contrast, the Matsuno–Gill pattern weakens over the TIO, without the development of the warm Kelvin wave in the equatorial western Pacific and anticyclonic anomalies over the NWP. Over the central Pacific, a pair of warm anomaly maxima develops in tropospheric temperature on either side of the equator. The cause of these off-equatorial maxima in tropospheric warming needs further investigation.

These results from the model hindcast indicate that the increase in TIO SST variability strengthens its influence on NWP climate. They support the notion that the warm tropospheric Kelvin wave is a key element of this teleconnection.

## 5. Antisymmetric wind pattern in spring and early summer

What caused the increase in TIO SST variability in summer after the mid-1970s? Previous studies show that El Niño induces TIO warming through atmospheric bridge (Alexander et al. 2002; Schott et al. 2009). Ocean Rossby waves are important over the southwest TIO (Xie et al. 2002; Huang and Kinter 2002), and surface heat flux adjustments dominate the rest of the TIO (Klein et al. 1999; Lau and Nath 2003).

Figure 8 shows the observed correlations with the December–February (0) [DJF(0)] Niño-3.4 SST index of SST and 850-hPa wind velocity averaged zonally over the TIO (40°–100°E). El Niño–induced SST warming persists long over the TIO, with the correlation staying above 0.6 through JJA(1) during the POST epoch but falling below this level before June(1) in the PRE epoch. During the POST epoch, an equatorially antisymmetric wind pattern develops in March and persists through June (Kawamura et al. 2001; Wu et al. 2008). Over the North Indian Ocean (NIO), the associated easterly wind anomalies act to warm SST after the onset of the southwest monsoon by reducing the prevailing winds and surface evaporation (Du et al. 2009). Over the NIO, the easterly wind anomalies during May–July explain the peculiar SST warming that grows after El Niño has already dissipated. During the PRE epoch, the antisymmetric wind pattern does not appear in spring, and wind anomalies are weak afterward over the NIO, where SST anomalies decay rapidly (Fig. 10a).

The model reproduces observed wind anomalies quite well for the POST epoch (Fig. 9b), including the antisymmetric wind pattern during March–May(1) [spring; MAM(1)]. In both observations and the model, easterly anomalies appear in a broad equatorial band during January–March and then gradually move northward and persist over the NIO through June and July. The easterly anomalies during May–July sustain the NIO warming, and their successful simulation in our AGCM hindcast indicate that they are forced by SST anomalies—most likely those over the southwest TIO (Du et al. 2009). Indeed, positive precipitation anomalies are found over the southwest TIO during both MAM(1) (Fig. 10b) and JJA(1) (Fig. 7b). The zonal-mean wind anomalies in the model do not correlate very well with observations for the PRE epoch (Fig. 9a)—the model winds are even westerly over the NIO during JJA(1).

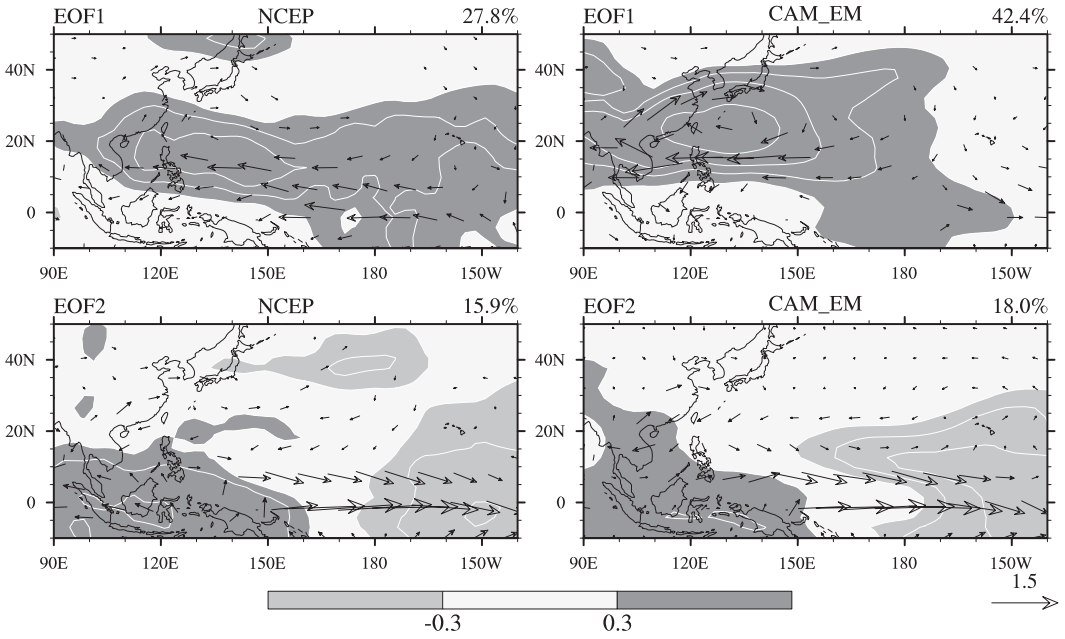



FIG. 4. Correlation of 850-hPa wind vectors and sea surface pressure in JJA with the leading PCs of 850-hPa wind velocity in (left) observations and (right) the ensemble-mean simulation: (top) EOF-1 and (bottom) EOF-2. The domain for EOF analysis covers the NWP (5°–50°N, 90°–180°E).

Figure 10 compares the leading EOF mode for MAM 850-hPa wind velocity and associated precipitation anomalies between the PRE and POST epochs in the model. As in summer, the model skills in MAM wind simulation are low before but high after the 1970s shift (the PC-1 correlation between the hindcast and observations changes from −0.1 to 0.68). During the POST epoch, EOF-1 explains a large fraction (42%) of the total variance, featuring an antisymmetric wind pattern with anomalous northeasterlies over the NIO. The associated precipitation anomalies are rough out of phase north and south of the equator, consistent with the result of Wu et al. (2008) that the leading precipitation EOF for spring is antisymmetric about the equator. The reduced rainfall north of the equator is accompanied by anticyclonic circulation, with easterlies over the NIO. A blob of positive rainfall anomalies is found over the southwest TIO, accompanied by cyclonic circulation. Over the southwest TIO, the slow propagation of downwelling ocean Rossby waves induces SST warming and intensifies atmospheric convection, thereby forcing the antisymmetric wind pattern across the equator (Xie et al. 2002; Annamalai et al. 2005; Du et al. 2009).

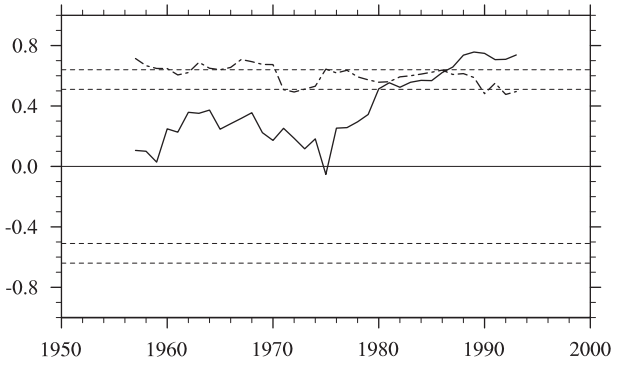



FIG. 5. Correlation between observations and the ensemble-mean simulation in 15-yr sliding windows for the first (solid line) and second (dash) PCs of JJA 850-hPa wind over the NWP. Horizontal black dash lines denote 95% and 99% confidence levels based on *t* test.

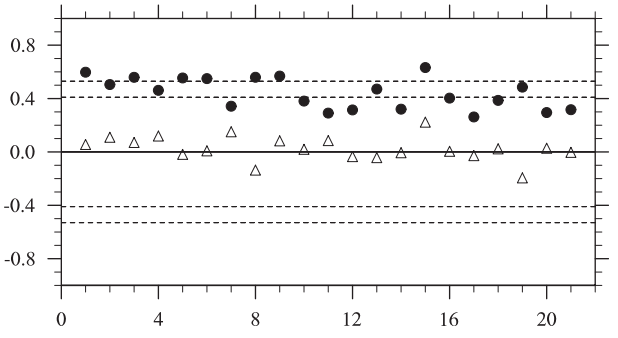



FIG. 6. Correlation coefficients between the JJA TIO SST and NWP anticyclone indices in the 21-member simulations during 1977–2000 (black circles) and 1953–76 (triangles). Horizontal dash lines denote 95% and 99% confidence levels.

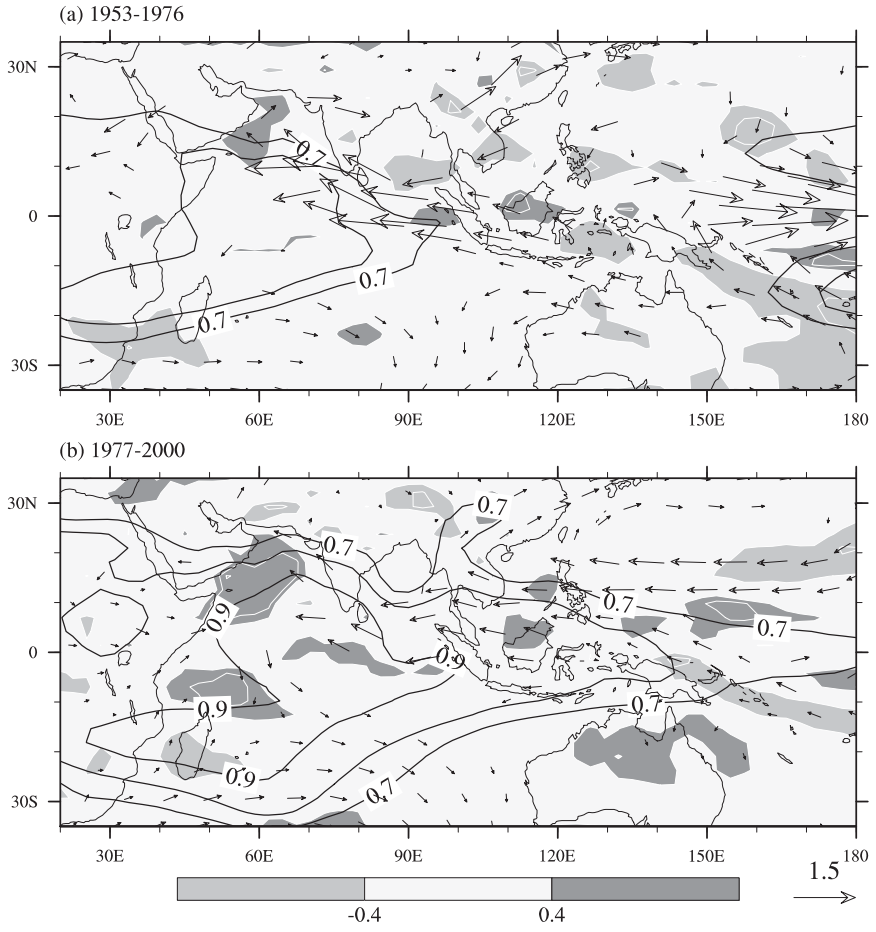



FIG. 7. As in Fig. 2, but for correlation with the JJA TIO index and ensemble-mean precipitation (gray shade and white contours at intervals of 0.2).

For the PRE epoch, EOF-1 explains a smaller fraction (33%) of variance and features easterly wind anomalies centered on the equator (Fig. 10a). The wind anomalies are consistent with a slowdown of the Walker circulation and resemble the wind pattern forced by remnant SST anomalies of a decaying El Niño (Annamalai et al. 2005). Precipitation anomalies are weak over the southwest TIO, consistent with the notion that they anchor the antisymmetric wind pattern across the equator.

Figure 11 shows the evolution of SST correlation with the model PC-1 for ensemble-mean MAM wind, to identify the SST forcing. From December to May, SST warming takes place over the entire TIO for the PRE epoch but is weak over the NIO for the POST epoch. Wu et al. (2008) suggest that the meridional SST gradient is important for the antisymmetric wind pattern, in line with wind–evaporation–SST (WES) feedback (Xie and Philander 1994). The TIO warming decays rapidly from June onward for the PRE epoch but persists through JJA for the POST epoch.

The TIO wind EOF-1 pattern for MAM is associated with Pacific SST anomalies indicative of El Niño. The evolution of Pacific SST correlation, however, is very different between the PRE and POST epochs. El Niño decays earlier during the PRE than the POST epoch. During the POST epoch, the slower decay of El Niño applies teleconnective forcing longer on the TIO, making the southwest TIO warming stronger and persist longer (Xie et al. 2010). The stronger SST forcing over the southwest TIO helps induce and persist the antisymmetric wind pattern, causing the NIO to warm during May–July. Positive SST anomalies over the southwest TIO have another effect of delaying the onset of the Indian summer monsoon, which may help sustain the NIO warming with increased solar radiation and reduced wind (Joseph et al. 1994; Annamalai et al. 2005).

For the POST epoch, we note a well-organized band of negative SST correlation over the subtropical NWP from December to May, conducive to forming a surface anticyclone (Wang et al. 2003). The NWP anticyclone

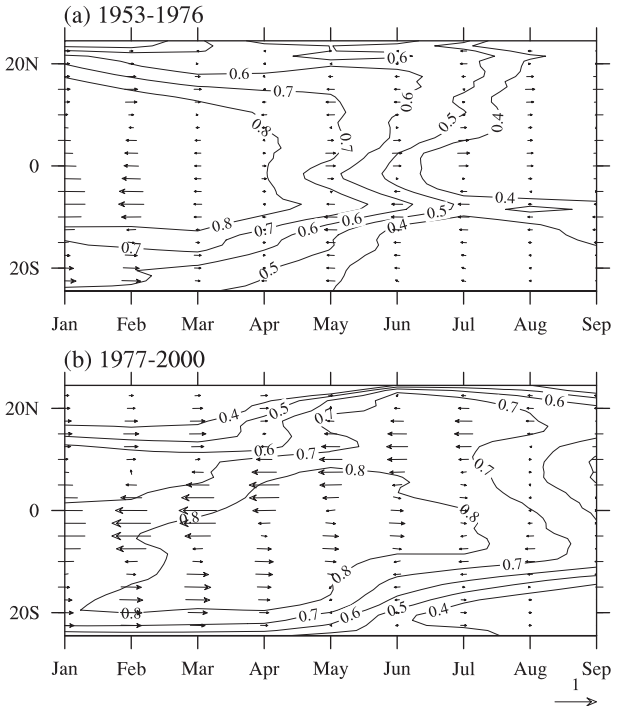



FIG. 8. Observed correlation (contours) between zonal-mean TIO SST (40°–100°E) and the DJF Niño-3.4 index, and regression (vector) of zonal-mean 850-hPa wind upon the Niño-3.4 index during (a) 1953–76 and (b) 1977–2000, as a function of calendar month and latitude.

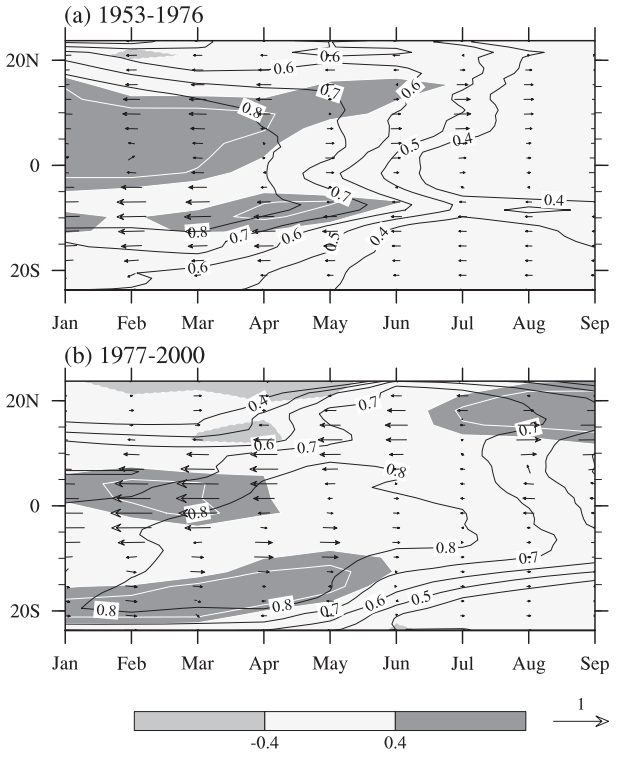



FIG. 9. As in Fig. 8, but for the ensemble-mean simulation. Superimposed is the correlation between zonal-mean (40°–100°E) precipitation (gray shade and white contours at intervals of 0.2) and the Niño-3.4 index.

may propagate westward into the NIO as an atmospheric Rossby wave, helping to suppress atmospheric convection there (Kawamura et al. 2001). This suggests that the NWP cooling may serve as a remote forcing for the antisymmetric wind pattern over the TIO.

## 6. Summary and discussion

The present study has investigated interdecadal change in the relationship between TIO SST and the NWP summer anticyclone using both observations and AGCM simulations. Observations show that the NWP anticyclone is negatively correlated with summer TIO SST after the mid-1970s, but the relationship is weak for the period of 1958–76. A 21-member ensemble AGCM hindcast reproduces this interdecadal change. Together, our observational and model results show that the TIO teleconnection to the NWP strengthens after the mid-1970s, most likely as a result of enhanced variability in summer TIO SST.

Our AGCM hindcast confirms that the interannual TIO warming excites a warm tropospheric Kelvin wave and that this Kelvin wave is instrumental in inducing summer climate anomalies over the subtropical NWP, including suppressed convection and an anomalous surface anticyclone. The model captures the pronounced Kelvin wave response of the troposphere to the El Niño–induced TIO warming for the POST epoch, but the Kelvin wave response is weak for the PRE epoch, explaining weak correlation between TIO SST and NWP climate anomalies. The strong (weak) TIO teleconnection coincides with high (low) SST variability over the summer TIO on one hand and with the period when the model skills are high (low) in simulating NWP circulation anomalies on the other.

An antisymmetric wind pattern develops in spring and persists through June, likely anchored by the positive SST and precipitation anomalies in the southwest TIO. With northeasterlies north of the equator opposing the mean southwest monsoon, this wind pattern is instrumental in the summer peak of the NIO warming (Du et al. 2009). This antisymmetric spring wind pattern is captured for the POST epoch but absent in the PRE epoch in the model. The SST correlation analysis suggests that stronger meridional SST gradients during the POST epoch rather than the PRE epoch contributes to the formation of this wind pattern, and so does the slower decay of El Niño SST anomalies in the Pacific.

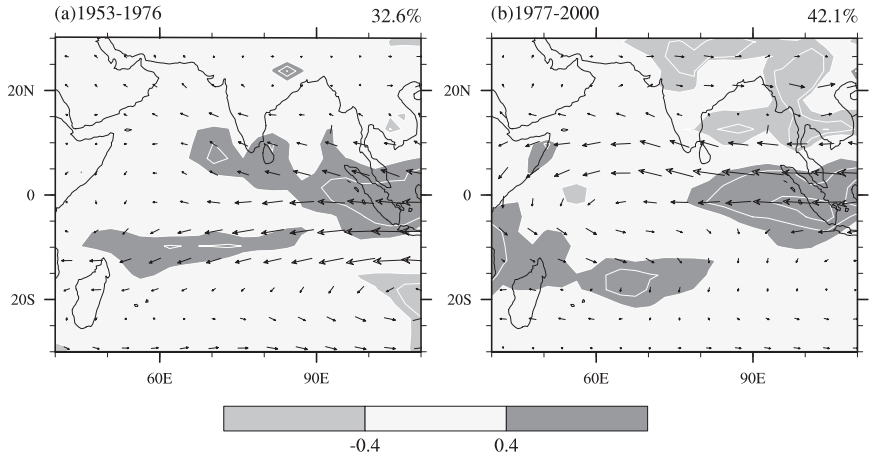



FIG. 10. First EOF modes of MAM 850-hPa wind velocity (vectors) over the TIO during (a) 1953–76 and (b) 1977–2000, with MAM precipitation correlation superimposed (gray shades and white contours at intervals of 0.2). The results are based on the ensemble-mean simulation.

Our AGCM hindcast prescribes observed SST, but SST anomalies could be interactive with wind anomalies. For example, the easterly wind anomalies and SST warming over the NIO may interact with each other: The NIO warming induced the NWP anticyclone and easterly anomalies over the Bay of Bengal–South China Sea as simulated in our hindcast (Fig. 7b) as these wind anomalies cause the ocean to warm (Du et al. 2009).

Results from our AGCM hindcast show that the interdecadal strengthening of both the antisymmetric spring wind pattern over the TIO and summer TIO teleconnection is due to changes in the pattern and magnitude of TIO SST anomalies. Such changes in TIO SST anomalies, in turn, appear to trace back to those in ENSO in the Pacific. TIO SST anomalies are interactive with the atmosphere and with Pacific SST via atmospheric bridge [see Schott et al. (2009) for a recent review]. The interdecadal change in ENSO and its interaction with the TIO needs to be studied from a coupled ocean–atmospheric perspective.

Two strong El Niño events happened after the mid-1970s: 1982/83 and 1997/98. We have computed the correlation coefficient between the TIO SST and NWP summer anticyclone indices during 1977–2000 by excluding the years of 1983 and 1998. The correlation coefficients drop to 0.49 and 0.41 for observations and the

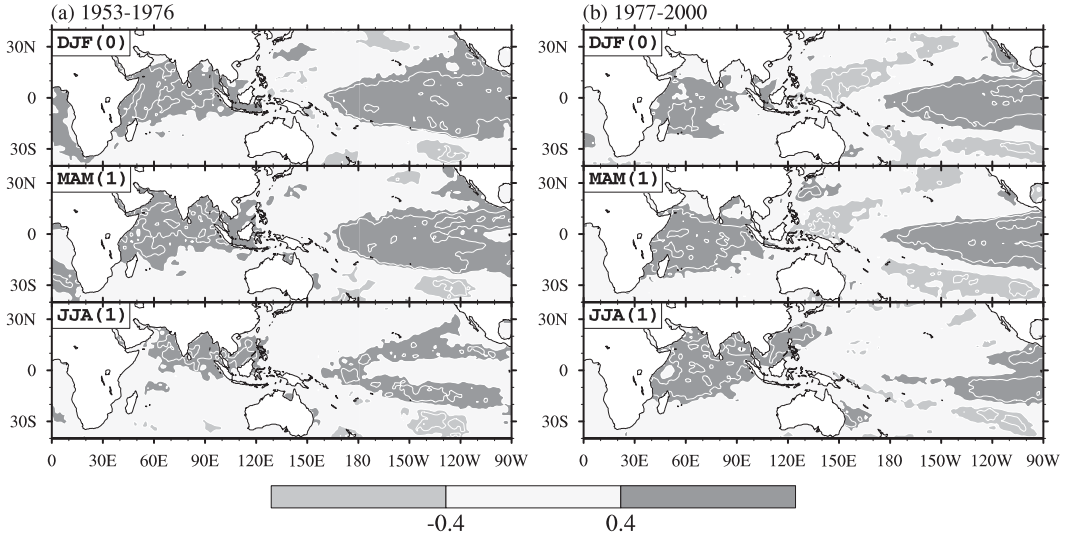



FIG. 11. Correlation of SST (gray shades and white contours at intervals of 0.2) with the first PC of MAM 850-hPa wind over the TIO from DJF to JJA during (a) 1953–76 and (b) 1977–2000.

model ensemble mean, respectively. They are lower than using the original time series (0.63 and 0.55 for observations and the simulation, respectively) but are still much higher than the correlation for the PRE epoch (Fig. 2). Thus, individual El Niño events can influence the relationship between the NWP summer monsoon and the Indian Ocean to some extent, but the sharp interdecadal change remains without the 1982/83 and 1997/98 events. The result does illustrate, however, the need to study low-frequency variations in ENSO properties.

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ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 28, NO. 2, 2011, 421–432

# The Impact of the Tropical Indian Ocean on South Asian High in Boreal Summer

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**ABSTRACT**

The tropical Indian Ocean (TIO) is warmer than normal during the summer when or after the El Niño decays. The present study investigates the impact of TIO SST on the South Asian High (SAH) in summer. When the TIO is warmer, the SAH strengthens and its center shifts southward. It is found that the variations in the SAH cannot be accounted for by the precipitation anomaly. A possible mechanism is proposed to explain the connection between the TIO and SAH: warmer SST in the TIO changes the equivalent potential temperature (EPT) in the atmospheric boundary layer (ABL), alters the temperature profile of the moist atmosphere, warms the troposphere, which produces significant positive height anomaly over South Asia and modifies the SAH.

An atmospheric general circulation model, ECHAM5, which has a reasonable prediction skill in the TIO and South Asia, was selected to test the effects of TIO SST on the SAH. The experiment with idealized heating over the TIO reproduced the response of the SAH to TIO warming. The results suggest that the TIO-induced EPT change in the ABL can account for the variations in the SAH.

**Key words**: tropical Indian Ocean, South Asia High, equivalent potential temperature

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## 1. Introduction

The ENSO is a major mode of air–sea interaction on interannual timescale with profound influences on global climate. The tropical Indian Ocean (TIO) gradually warms when El Niño is in its developing phase, and the warming reaches a maximum nearly one season after the NINO3.4 SST anomaly has peaked (Klein et al., 1999; Lau and Nath, 2000; Alexander et al., 2002; Lau and Nath, 2003). The warm anomaly still exists during the summer after the peak of El Niño (Du et al., 2009). Klein et al. (1999) pointed out that changes in evaporation and cloud cover induced by the El Niño account for the warm anomaly over most parts of the Indian Ocean during the ENSO decaying phase; while the warm anomaly over the southwest Indian Ocean could not be explained by the above changes. Previous research (Masumoto and Meyers, 1998; Yu et al., 2005) revealed that, when the ENSO matures, changes in the Walker circulation generate an anticyclonic wind anomaly over the TIO, which forces downwelling Rossby waves in the southeast Indian Ocean. The Rossby waves propagate westward, arrive in the southwest Indian Ocean several months later, deepen the thermocline ridge and warm the SST (Huang and Kinter, 2002; Xie et al., 2002). Thus, ocean dynam-

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ics may play an important role in the warming in the southwest Indian Ocean.

The warming of the TIO in response to El Niño has influences on the climate locally and remotely after El Niño decays. This phenomenon is called the "capacitor effect" (Annamalai et al., 2005; Yang et al., 2007; Yang and Liu, 2008; Xie et al., 2009). In the years when there is basin-wide warming in the TIO, anticyclonic wind anomaly occurs over the western North Pacific, rainfall over Yangtze River is above normal and the South Asian High (SAH) strengthens in the summer (Yang et al., 2007). The Fast Ocean-Atmosphere Model and five Atmospheric General Circulation Models (AGCMs) have reproduced the phenomena (Yang et al., 2007; Li et al., 2008). Analysis by Xie et al. (2009) showed that a Kelvin wave anomaly induced by warming in TIO is key to the generation of an anticyclone anomaly over the western North Pacific. Earli er studies (Wang et al., 2000, 2003) suggested a role played by local air-sea interaction in the maintenance of the western North Pacific anomalous anticyclone in early summer.

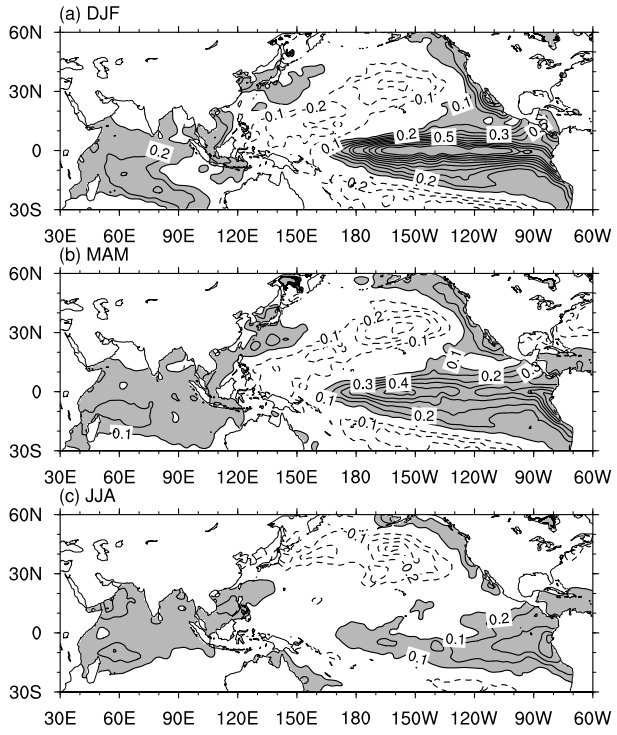



**Fig. 1.** Regression of SST against N(0)D(0)J NINO3.4 SST index from 1979–2007, (a), (b) and (c) are the results of D(0)JF, MAM and JJA, respectively. Negative values are shown in dash contours. Shading indicates values larger than 0.1; and the interval of white contours is 0.1 K, the unit is K. For the sake of clarity, spatial 9-point smoothing is performed and the zero contour is omitted.

Figures 1 and 2 demonstrate the regression of SST and 200-hPa velocity potential, respectively, on the NINO3.4 SST index from November(0) to January (N(0)D(0)J). The numeral "0" denote the preceding year. In the winter when El Niño peaks and the following spring, SST is warmer than normal in the equatorial eastern Pacific (EEP) and the 200-hPa velocity potential shows an upper-level anomalous divergence. Though there is a warming in the TIO, the 200-hPa velocity potential suggests a weak anomalous divergence or convergence over the TIO. This indicates that during these two seasons, the EEP SST has a major impact on circulation over the tropical Indo-Pacific Ocean and that TIO SST does not have a large impact on circulation or it just responds to the atmosphere passively. During the summer following El Niño, the warm anomaly over the EEP is weaker than that in N(0)D(0)J and March to May (MAM), and the intensity of the warm anomaly over the TIO is approximately equal to that in N(0)D(0)J and MAM. The 200-hPa velocity potential in summer shows an anomalous divergence over the TIO and no significant anomaly over the EEP, indicating an active role of the TIO SST on local and surrounding circulation in sum-

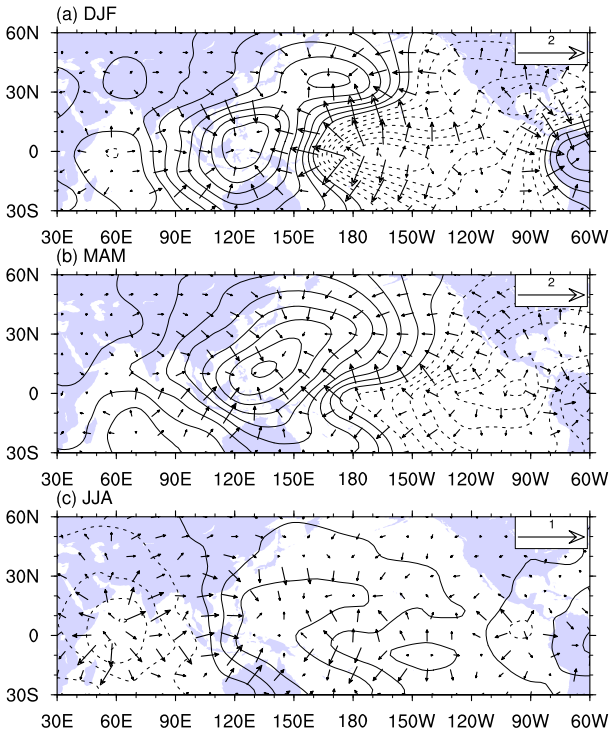



**Fig. 2.** As in Fig. 1, except for the results of 200-hPa velocity potential (contours) and divergent wind component (vectors), the contour interval is $2.5\times10^5$ m$^2$ s$^{-1}$. Solid and dashed lines represent positive and negative values, respectively.

mer, in agreement with Wu et al. (2009). The regression of SST and 200-hPa velocity potential in the preceding winter, spring, and summer, on the TIO SST in summer shows similar results. The present study focuses on the impacts of the TIO SST in summer.

The SAH is a high pressure system located in the upper troposphere over the Tibetan Plateau and its surrounding area in summer. It is a result of the elevated heating of the Tibetan Plateau to the atmosphere (Flohn, 1960) and has been recognized to be the strongest and steadiest system in the upper troposphere (Mason and Anderson, 1963). The variations of the SAH are tightly linked to precipitation and circulation over Asia. Tao and Zhu (1964) reported that longitudinal displacement of the SAH leads that of a 500-hPa subtropical anticyclone over the Western Pacific for a few days and is a good index for short- to medium-range weather forecasts in Asia. In the summer, when the SAH shifts eastward, significant positive rainfall anomalies occur in southern Japan, over the Korea Peninsula, and in the Yangtze River-Yellow River Valley of China and the Tibetan Plateau. As the SAH shifts westward, the rainfall over these areas tends to be below average (Zhang et al., 2002). Zhang and Wu (2001) pointed out there is a good relationship between the strength of the SAH and rainfall over the Yangtze River valley in summer. The variation of the summer SAH can modulate climate downstream. A strengthened SAH emanates anomalous wave energy downstream along the westerly jet stream, leading to the occurrence of an intensified subtropical high over the extratropical North Pacific (Zhao et al., 2009). Also, the summer SAH is linked to global circulation and precipitation. A strengthened SAH is accompanied by a strengthened and westward Western Pacific subtropical high, weakened mid-Pacific trough and strengthened Mexican high. Furthermore, rainfall tends to increase over South Asia, Central America, Australia and Central Africa, and decrease over the Pacific and Mediterranean (Zhang et al., 2005). In addition, the easterly and westerly jets residing on the south and north sides of the SAH form a transport barrier of some chemicals, such as water vapor and ozone (Dethof et al., 1999; Randel and Park, 2006). Meanwhile, because of upward transportation in the center of SAH, some chemicals such as carbon monoxide are transported to the upper troposphere and trapped there (Park et al., 2004; Li et al., 2005).

Because of the large heat content of water, warm oceans heat the atmosphere and exert deep influence on circulation. The TIO and SAH are adjacent in geographical position, and change in thermal regime over the TIO may affect the variability of the SAH. The Indian Ocean basin-wide mode strengthens the summer SAH probably through the Matsunuo-Gill response forced by condensational heating associated with increased precipitation over the TIO (Yang et al., 2007; Yang and Liu, 2008; Li et al., 2008).

In addition to changes in its strength, the SAH also varies in its latitudinal and longitudinal locations (Zhang et al., 2000, 2002). Although there have been studies on the relationship between the TIO SST and SAH strength, the impact of the TIO SST on the displacements of the SAH is not clear. At the same time, former studies on the effects of the TIO SST on the SAH do not clearly address the influencing mechanism. Our present analysis will show that precipitation over the TIO is not significantly correlated with TIO SST. The influence mechanism is therefore still not well understood, an issue that will be discuss further, below.

The paper is organized as follows. Section 2 describes data, the method and AGCM experiments. Section 3 analyses the effects of TIO SST on SAH variations and the possible influencing mechanism. Section 4 uses simulations to validate the relationship and influencing mechanism proposed in section 3. Section 5 presents a discussion and summary of the findings.

## 2. Data and method

### 2.1 *Observational data*

This study used the Hadley Center Global Sea Surface Temperature (HadISST) dataset (Rayner et al., 2006), the Center for Climate Prediction (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997), and the National Centers for Environmental Prediction–National Center for Atmosphere Research (NCEP–NCAR) atmospheric reanalysis dataset (Kalnay et al., 1996). The horizontal grid resolutions of these datasets are $1^\circ\times1^\circ$, $2.5^\circ\times2.5^\circ$, and $2.5^\circ\times2.5^\circ$, respectively. In order to be consistent with the period covered by the CMAP dataset which starts from 1979, and considering that there is interdecadal change in the influence of TIO SST during the late 1970s (Xie et al., 2010), the analysis and discussion in this paper will focus on the period 1979–2007.

Another focuses of the study is on the interannual variability of the TIO and SAH. For this purpose, three-month mean data were obtained and used in order to remove intraseasonal variability. Further, decadal and longer variations, which are significant over the TIO and surrounding areas (Du and Xie, 2008), were removed.

Correlation, regression and composite analysis were used to examine the relationship between the two variables. The statistical significance of a composite difference was assessed using the Student's $t$ test. All

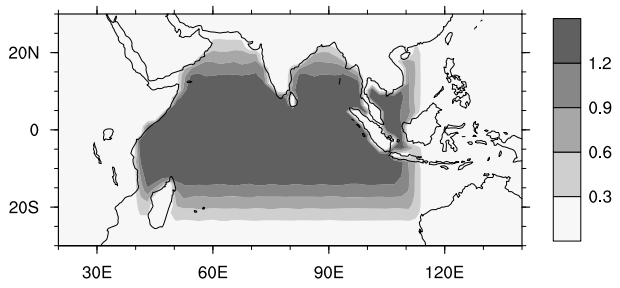



**Fig. 3.** Idealized SSTA imposed on climatological SST in ECHAM5.

significances referred to in this paper are at the 95% confidence level, unless otherwise stated. Finally, summer, in the context of this study, refers to the period from June to August.

### 2.2 *AGCM experiment*

The AGCM was ECHAM5, the latest Hamburg version of the European Center for Medium-Range Weather Forecast model, with a resolution of T63L19 (Roeckner et al., 2003).

Two experiments were designed. The first used prescribed SST and sea-ice from 1957–1999 to drive ECHAM5 (hereafter, AMIP2). The results of this experiment will be used to examine the ability of the AGCM with respect to SAH and climate over the TIO and the surrounding regions. In order to be consistent with the period of observational analysis, a dataset with covering the period 1979–1999 was selected. In the second experiment, referred to as the TIO sensitivity experiment (TIOSE), the AGCM was driven by an idealized SST anomaly in the TIO superposed on the model's climatological monthly SST. The SST anomaly is maximum along the equator and gradually decreases to 0 at 30°N and 30°S (Fig. 3). The SST anomaly increases linearly at a rate of 0.2 K $yr^{-1}$ from 0 to 2.2 K. Five simulations were performed with different initial conditions. In the discussion, the ensemble mean is analyzed to remove the uncertainties induced by the initial conditions and internal variability.

## 3. observations

### 3.1 *Relationship between TIO SST and the SAH*

Figure 4a illustrates normalized June–August (JJA) TIO (40°–100°N, 20°S–20°E) SST time series. Seven warm cases (1983, 1987, 1988, 1991, 1998, 2003, and 2007) and nine cold cases (1981, 1984, 1985, 1989, 1994, 1996, 1999, 2000, and 2004) were chosen when the JJA TIO SST index exceeded 0.5 and −0.5 standard deviation, respectively.

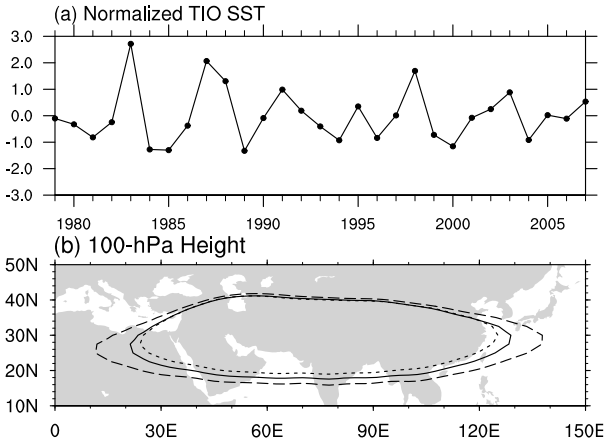



**Fig. 4.** (a) Normalized TIO (20°S–20°N, 40°–100°E) SST in June–August from 1979–2007. (b) Contour lines for 100-hPa 16720 gpm geopotential height when TIO SST is warm (dashed line), cold (dotted line) and climatological (solid line) in JJA.

Figure 4b shows contour lines for 16720 gpm of 100-hPa geopotential height when TIO SST represents warm cases (dashed contour line), cold cases (dotted contour line) and climatology (solid contour line). When the TIO is warmer (or colder) than normal, the SAH strengthens (or weakens) and moves southward (or northward). As TIO SST moves from cold to warm, the center of the SAH shifts southward, its southern boundary moves southward, the eastern and western boundaries respectively shifts eastward and westward, while almost no movement is found in the northern boundary. The definition of the SAH indices follow those in Zhang et al. (2000) and Zhou et al. (2006). The center of the SAH is the grid where the geopotential height is the largest and is expressed by its latitude and longitude. The area of the SAH is the total number of the points where the geopotential height is no less than 16600 m in the region (10°S–50°N, 30°W–180°E). The SAH intensity is the sum of differences between 16600 m and the geopotential height at each point where the geopotential height is no less than 16600 m in that region. The correlation coefficients between the TIO SST and SAH indices (including strengthening index, area index, center latitude and center longitude), are 0.63, 0.53, −0.43 and −0.08, respectively. Among the four correlation coefficients, the first two reach the 99% confidence level and the third reaches the 95% significance level. Thus when the TIO is warmer, not only does the SAH strengthen, but its center and main body also shift southward. However, at the interannual timescale, no direct relationship could be found between TIO SST and zonal displacement of the SAH in boreal summer.

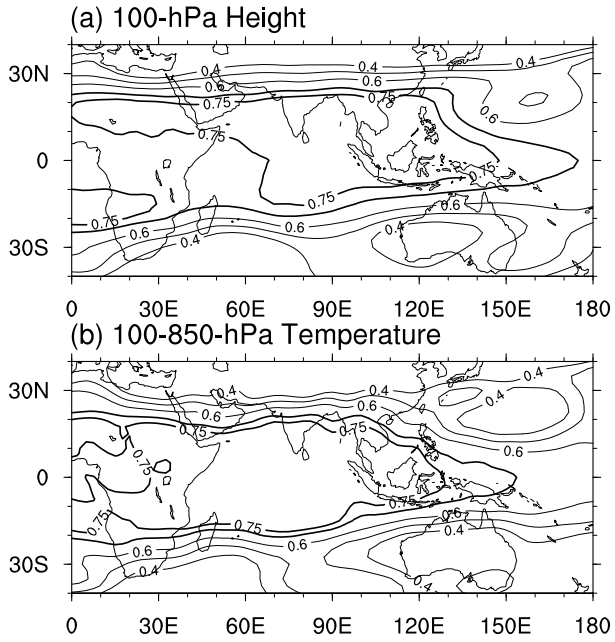



**Fig. 5.** Correlation with JJA TIO SST index: (a) 100-hPa geopotential height, (b) tropospheric temperature. The value 0.4, 0.5, 0.6, 0.65, 0.7 and 0.75 are shown in contours, and the value 0.7 and 0.75 are shown in thick contours.

### 3.2 *Possible mechanism of impact*

Figure 5 shows the correlation distribution of 100-hPa geopotential height (hereafter H100) and tropospheric temperature (100–850-hPa temperature; hereafter TT) with TIO SST. In the H100 field (Fig. 5a), a significant positive anomaly is situated in the tropical and subtropical region over the TIO and surrounding areas. The height anomaly over the domain (15°–30°N, 30°–120°E), south of the SAH's climatological location, is the key area for the strengthening and southward displacement of the SAH in boreal summer. As anomalies of H100 are closely related to the variation in 100–850-hPa temperature, TT field was analyzed. In the TT field (Fig. 5b), to the west of the Indian Ocean, two maximums reside off the equator over tropical Africa; to the east, the maximum lies along the equator over the maritime continents. The TT anomaly over the Indian Ocean and the surrounding regions exhibits a Matsunuo-Gill pattern in response to local heating (thick contours in Fig. 5b), indicating an active area of the TIO on the atmosphere.

Previous studies (Yang et al., 2007; Yang and Liu, 2008) on the impacts of the TIO on the SAH have suggested that condensational heating associated with precipitation anomalies plays an important role in the effects of the TIO on the SAH. However, the correlation results between TIO SST and precipitation do

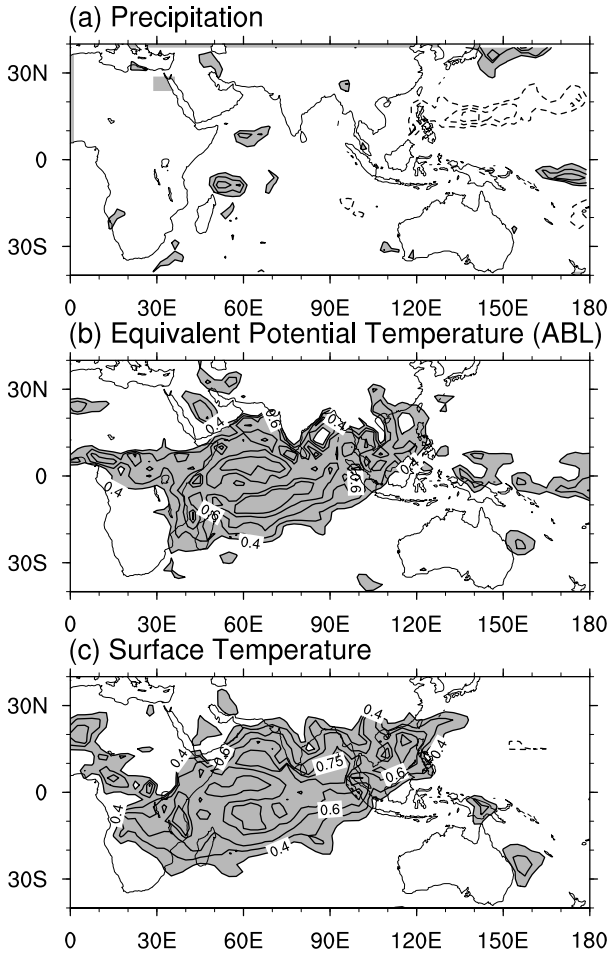



**Fig. 6.** Correlation with JJA TIO SST index: (a) precipitation; (b) EPT in 1000-hPa and (c) surface temperature. Shading indicates values larger than 0.4. The value −0.75, −0.7, −0.6, −0.5, −0.4, 0.4, 0.5, 0.6, 0.65, 0.7 and 0.75 are shown in contours; negative values are shown in dash contours.

not exhibit large scale anomalous rainfall, and significant precipitation anomalies only locate over the southwest Indian Ocean and south Arabian Sea (Fig. 6a). The weak correlation between TIO SST and precipitation in boreal summer has also been pointed out by Xie et al. (2009). For convenience, the area average of H100 in summer over the domain (15°–30°N, 30°–120°E), which is the key area for the strengthening and southward displacement of the SAH, is defined as an index, named SAHSE; also, the area mean of precipitation over the domain (20°S–20°N, 40°–100°E) is defined as an index, named PTIO. The correlation coefficient between SAHSE and PTIO is 0.25, not reaching the 90% significance level. Thus, the correlation analysis indicates that condensational heating associated with local precipitation over the Indian Ocean does not play the key role in the response of the H100.

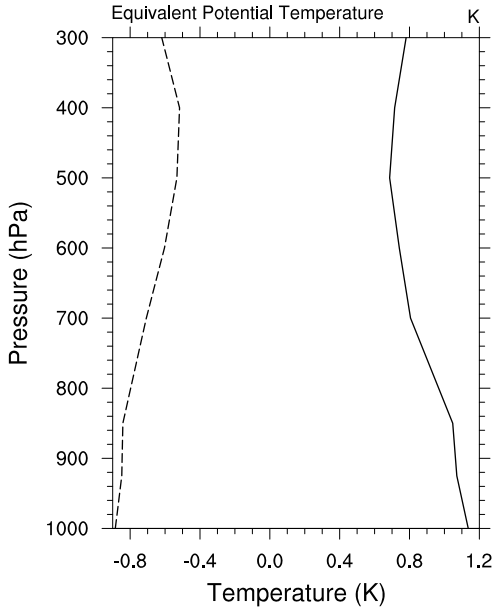



**Fig. 7.** Anomalous EPT (K) profile over the TIO when the TIO is warm (solid line) and cold (dash line).

In order to better understand the tropospheric warming over the TIO when the TIO is warm, the theory of moist adjustment is introduced. In the tropical wet atmosphere, the temperature profile follows approximately a moist-adiabatic profile which can be determined by moist static energy (MSE) or equivalent potential temperature (EPT) in the atmospheric boundary layer (ABL; Emanuel et al., 1994, 1997; Su and Neelin, 2003). It was pointed out by Xie et al. (2009) that the TIO warms the atmospheric column through modifying the EPT or MSE in the ABL. Figure 7 shows the anomalous EPT profiles when the TIO is warm and cold in boreal summer. As the specific humidity dataset in the NCEP-NCAR reanalysis is limited up to 300 hPa, the EPT profile was calculated between that level and 1000 hPa. Unlike the latent heat profile with a prominent peak at one level, the EPT profile shows an approximate uniform warm (or cold) profile from lower to upper level, further indicating that precipitation-induced latent heat may not play the key role. Correlation of EPT1000 with TIO SST shows a large-scale significant positive anomaly over the TIO (Fig. 6b). The correlation coefficient between the area average of EPT1000 over the domain (20°S–20°N, 40°–100°E) and SAHSE is 0.75, far exceeding that between PTIO and SAHSE. The correlation coefficient reveals that TIO-SST-induced EPT change in the ABL is a more important manner through which TIO SST effects tropospheric warming.

In brief, in boreal summer, the warming of the TIO induces EPT variation in the ABL, modulates the temperature profile of the atmosphere above it, heats the atmospheric column, elevates H100 with a significant positive H100 anomaly to the south of the SAH's climatological location, strengthens the SAH and leads to a southward shift of its center.

Besides the direct influence of the TIO on the SAH, some indirect processes, such as TIO-SST-anomaly-induced anomalous conditions on land and the thermal contrast between land and ocean, may also contribute to variation in the SAH. When the TIO is warm in boreal summer, precipitation increases over the north of the Bay of Bengal (Fig. 6a), in accordance with the simulation results of Yang and Liu (2008). The associated condensational heating contributes to tropospheric warming over the area and forces an anomalous anticyclone in the upper troposphere over the SAH's climatological position. These processes are favorable for the strengthening and southward displacement of the SAH. As the TIO warms, warm anomalies in surface temperature are found in the southwest of the Indian Peninsula, east of the Indian and north of the Indo-China Peninsula (Fig. 6c). It can be inferred that the consequent sensible heat anomaly may also contribute to tropospheric warming. Moreover, TIO warming may reduce the thermal contrast between the Indian Ocean and the continent of Asia, thus height difference over the two regions is reduced and a positive height anomaly occurs over the TIO region. This may also cause a strengthening and southward displacement of the SAH.

## 4. Model results

AGCM experiments were designed to test the hypothesis that TIO SST affects the strength and latitudinal displacement of the SAH in summer. For convenience, NCEP and CMAP results are referred to as "observation", though the NCEP reanalysis is obviously not observational data.

### 4.1 *Prediction ability of ECHAM5*

In the observation, summer mean precipitation in climatology locates mainly in the equatorial Indian Ocean, east Arabian Sea, Indian Peninsula and the Bay of Bengal, with a maximum over the equatorial southeast Indian Ocean, west of the Indian Peninsula and northeast of the Bay of Bengal (Fig. 8a). For the summer mean TT field, a ridge exists from northern Madagascar to the Somali Peninsula and two maximums exist over the Iran and Tibet Plateaus (Fig. 8a). In H100, a large-scale high system, the SAH, locates over Iran and Tibet Plateaus, with its center above the Tibet Plateau (Fig. 9a).

In ECHAM5, the large-scale characteristics of cli-

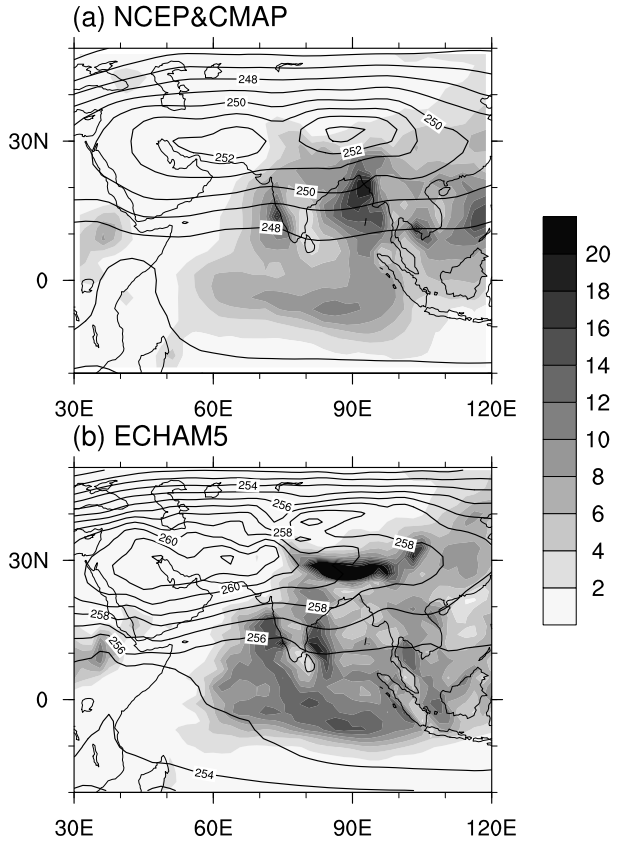



**Fig. 8.** Climatology of tropospheric temperature (contours) and precipitation (shaded) in JJA, (a) results of NCEP and CMAP; (b) results of ECHAM5. The units of tropospheric temperature and precipitation are K and mm $d^{-1}$, respectively.

matological precipitation in summer are consistent with those in observation, with large-scale precipitation along the equatorial south Indian Ocean, Arabian Sea and the Bay of Bengal; ECHAM5 also captures well the maximums in the equatorial southeast Indian Ocean and west Indian Peninsula, but fails to capture the maximum in the northeast of the Bay of Bengal (Fig. 8b). In ECHAM5, the TT ridge in the observation is reproduced well and there is also a maximum over 30°N. Although, the maximum shifts westward compared to the observation, and locates over the Iran Plateau (Fig. 8b). The value of simulated H100 is larger than that of the observation, and the SAH is reproduced but shifts westward relative to the observation (Fig. 9b). Therefore ECHAM5 can simulate tropospheric temperature, precipitation and geopotential height near the Indian Ocean well, and the climatology of the SAH is reproduced quite well. It is reasonable to discuss the impact of TIO SST on the strengthening and latitudinal variations of the SAH by using ECHAM5.

Owing to the difference of H100 value between ECHAM5 and the observation, 16860-gpm contour is adopted to characterize the SAH in simulation. Figure 10a illustrates the contour lines for 16860-gpm of 100-hPa geopotential height when JJA TIO SST is warm, cold and climatological. As the TIO warms (or cools), the SAH strengthens (or weakens) and shifts slightly southward (or northward), consistent with observed results.

### 4.2 *The TIO sensitivity experiment*

In order to investigate the effect of TIO SST alone and exclude the influences of SST over other oceans, TIOSE was analyzed. Figure 10b illustrates contour lines for 16860 gpm of 100-hPa geopotential height when the SSTA imposed on the TIO is 0–1 K and 1–2 K mean. Compared to the results of 0–1 K mean, the SAH strengthens and expands southward when the TIO SSTA is 1–2 K mean, with more significant movement in the southern boundary than that in the northern boundary. The SAH indices in this experiment were defined according to the definition in Zhang et al. (2000) and Zhou et al. (2006), except that the characteristic contour is 16 860 gpm. The scatter plots of SAH intensity (Fig. 11a), area (Fig. 11b) and center latitude (Fig. 11c) in TIOSE show good linear relation with SSTAs imposed on the TIO (the linear regression of all three reach the 99.9% significance level), while the linear relationship between center longitude and SSTA is poor (Fig. 11d). The results are well in accordance with the relationship between the TIO and SAH variation in the observation.

Figure 12 shows the differences in H100, TT, precipitation and EPT1000 between 1–2 K SSTA mean and 0–1 K SSTA mean. The H100 field (Fig. 12a) shows a significant positive height anomaly extending northward to 40°N. The height anomaly decreases from tropics to the mid latitude, leading to more significant movement in the SAH's southern boundary than in its northern boundary, as well as the southward displacement of its center. The response of TT shows two maximums off the equator, one to the west of the Indian Ocean and the other along the equator (Fig. 13b), which features a Matsuno-Gill type response. EPT1000 shows a large-scale significant positive anomaly over the TIO, also in good accordance with the observation (Fig. 12d). The results of H100, TT and EPT1000 support the notion that the TIO warms the troposphere through modulating EPT or MSE in the ABL, strengthens the SAH and makes it expand southward.

Compared to observations, the simulated precipitation is different, with a significant positive precipitation anomaly over the east Indian Ocean, and a

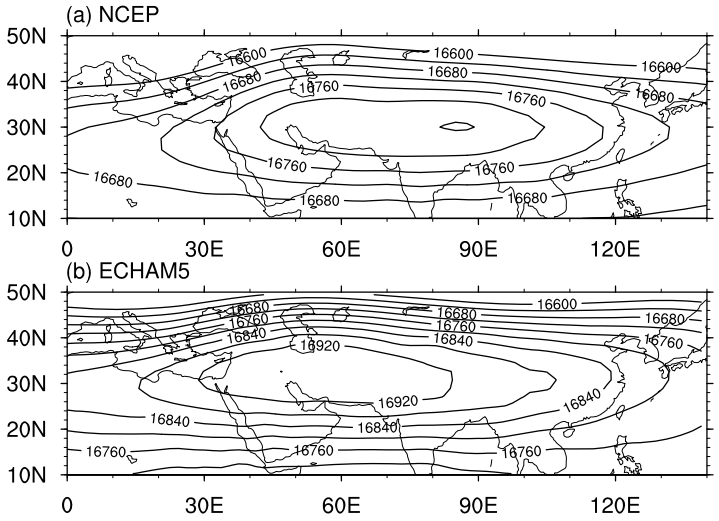



**Fig. 9.** Climatology of 100-hPa geopotential height (m) in JJA: (a) NCEP, (b) ECHAM5.

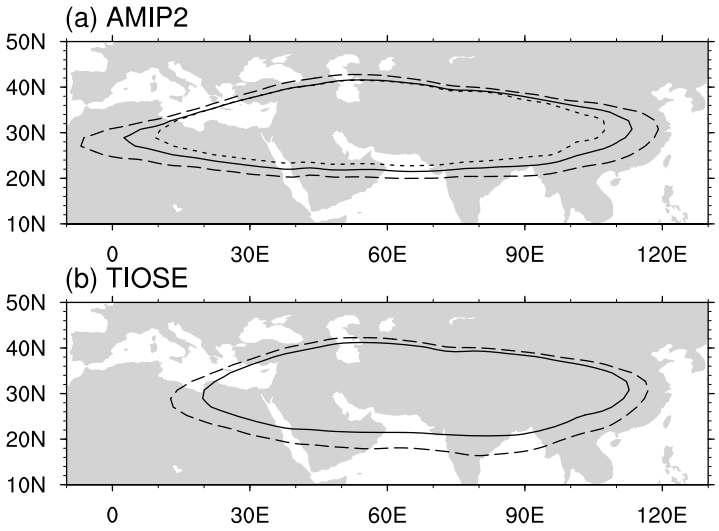



**Fig. 10.** (a) As Fig. 5, except for 16 860 gpm contour lines in AMIP2 run of ECHAM5. (b) Contour lines for 16 860 gpm of 100-hPa geopotential height when summer TIO SSTA is 0–1 K (solid contour lines) and 1–2 K (dashed contour lines).

negative anomaly over central North Africa from 10°–20°N, Tanzania, the Bay of Bengal and the maritime continents (Fig. 12c). Although the precipitation response over the Bay of Bengal in TIOSE is opposite to that in observations, the SAH strengthens and shifts southward in both TIOSE and the observations. So compared to the effects of precipitation over the Bay of Bengal, another process may be more important to the response of the SAH.

Su and Neelin (2003) pointed out that assuming remote SST is fixed, as local SST increases, moisture in atmospheric boundary adjusts quickly, the temperature profile of the atmosphere changes, convective available potential energy increases, and convective precipitation occurs, making the actual temperature profile close to a moist-adiabatic profile, meaning the atmosphere tends to be stable. Convection acts as a key factor in modulating actual temperature close to a moist-adiabatic profile. Thus there is a good relationship between precipitation and SSTAs under these condition. However, in the observation, when the TIO is warm in summer, a SSTA also exists over other oceans (Fig. 1). Provided the TIO SSTA is fixed, SSTAs outside the Indian Ocean may affect the

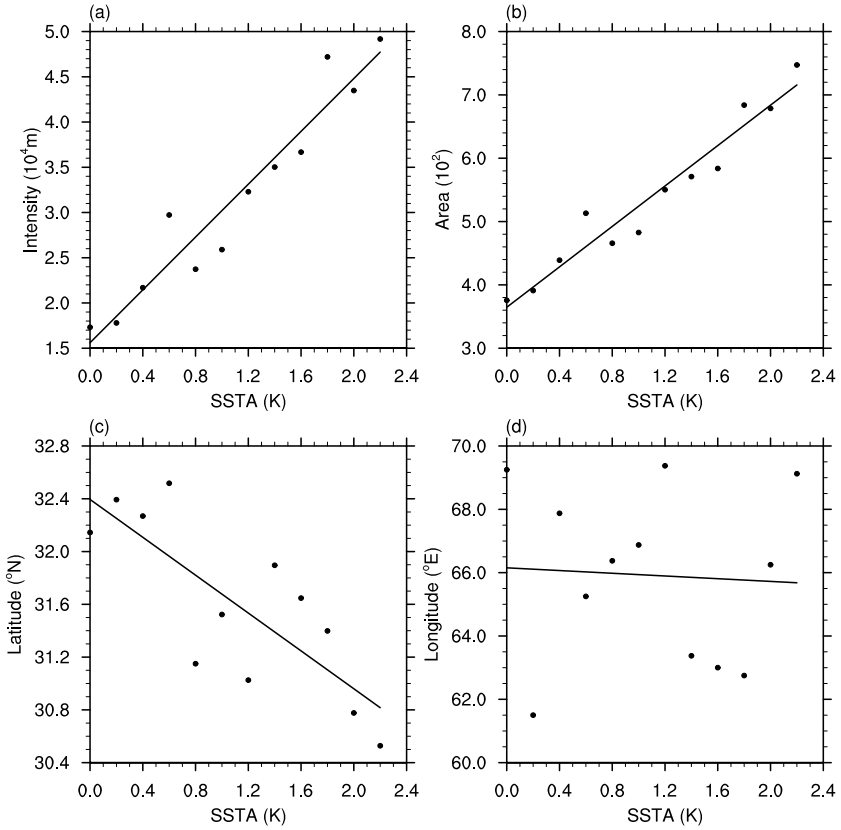



**Fig. 11.** Scatter plots of SAH (a) intensity ($\times 10^4$ m), (b) area ($\times 10^2$), (c) center latitude (°N) and (d) center longitude (°E) in JJA with respect to SSTA in TIOSE. Linear lines in the four plots are the linear fit lines.

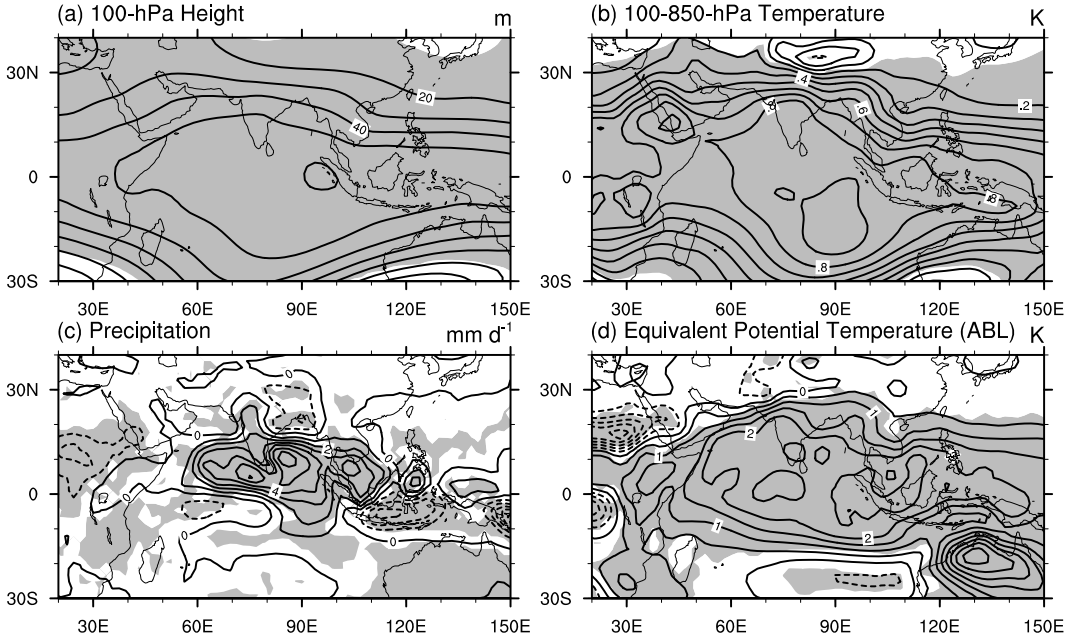



**Fig. 12.** Differences between 1–2 K mean and 0–1 K mean in TIOSE: (a) 100-hPa geopotential height (m); (b) tropospheric temperature (K); (c) precipitation (mm $d^{-1}$) and (d) EPT at 1000-hPa (K). Shading indicates the values reaching the 95% significance level. For the sake of clarity, spatial 9-point smoothing is performed.

atmosphere over the Indian Ocean, tending to make it warmer or colder. As the TIO SSTA is fixed, an anomalous EPT in the ABL over the TIO varies very little. From the changes in the vertical temperature profile, the atmosphere over the TIO tends to be more stable or unstable than normal and decreased or increased precipitation occurs to modulate the atmosphere to a neutral state. If the heating of the Indian Ocean is taken into consideration, a warm (or cold) TIO makes the actual temperature profile get close to a moisture-adiabatic profile through increased (or decreased) convective heating. Due to the combined effects of SSTAs over the TIO and outside the TIO, precipitation over the TIO is not significant. Thus the impacts of other oceans on TT over the TIO may account for the insignificant precipitation when the TIO is warm and it is not difficult to understand that precipitation does not necessarily correlate well to local SST. Despite that, the mechansim that the TIO warms the troposphere through modulating EPT or MSE in the ABL and affects the SAH can be applied to explain the results in both TIOSE and the observation. The heat budget of the troposphere over the TIO is still not understood and needs further investigation.

## 5. Discussion and summary

The effects of TIO SSTA on variations of the SAH and the possible underlying mechanism have been investigated. In the summer when the ENSO decays, the SST and circulation anomalies indicate that TIO SST directly or indirectly plays an active role in circulation changes over the TIO, with an active role by the eastern Pacific not being obvious. This phenomenon is not obvious in spring, nor the preceding winter. Thus the impact of TIO SST in summer was the focus of this study.

Compared to cold cases, when the TIO warms in summer, the main body of the SAH becomes larger, the SAH intensifies and its center shifts southward. A significant correlation can be found between TIO SST and the indices (intensity, area and center latitude) of the SAH, while the relationship between TIO SST and the SAH center longitude is poor. When the TIO warms, the whole troposphere over the TIO is heated with a Matsunuo-Gill response in the troposphere. No significant large-scale precipitation anomaly can be found over the TIO, indicating that precipitation does not act as a key factor in forcing the Matsunuo-Gill response. Emanual et al. (1994, 1997) and Su and Neelin (2003) pointed out that, the temperature in the tropical atmosphere follows approximately a moist-adiabatic profile which is determined by EPT or MSE in the ABL. TIO SST modulates tropospheric temperature through changing the EPT or MSE in the ABL, forcing the Matsunuo-Gill response. As the TIO warms, because of the spread of tropospheric warming and the Rossby wave response over the northern Indian Ocean, a positive height anomaly occurs in the south of the SAH's climatological location in summer, making the SAH intensify and its center move southward.

AGCM ECHAM5, with a resolution of T63L19, was used to test the effects of TIO SST on variations in the SAH. The model experiment indicated that, as the summer TIO SSTA increases, the SAH strengthens and its center shifts southward significantly. When the TIO warms, a significant tropospheric warm anomaly locates over tropics and subtropics, featuring a Matsuno-Gill type response over the TIO, consistent with the observation. However, the precipitation response to the TIO SSTA is different to that in the observation. By using the theory proposed by Su and Neelin (2003) and analyzing the results obtained in the present study, it can be inferred that when the TIO is warm, SST in other oceans may make a positive or negative contribution to precipitation over the TIO. Thus precipitation over the TIO is not all determined by local SST and the influences of SSTs in other oceans may account for the insignificant precipitation over the TIO in the observation. Although simulated precipitation was different from observed precipitation, the simulation results can still be explained by the mechanism that the TIO warms the troposphere through the modulation of the temperature profile and has impact on the SAH.

The simulations have shown that TIO warming contributes to the strengthening and southward displacement of the SAH in summer. However, the correlation coefficient between the SAH center latitude and simultaneous NINO3.4 index is $-0.45$, a little higher than that between TIO SST and the SAH center latitude. Therefore, besides the influence of the TIO, a simultaneous influence of eastern Pacific SST may also contribute to the latitudinal movement of the SAH in summer. How the former affects the latter also need further study.

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# The Interannual Variability of Summer Rainfall in the Arid and Semiarid Regions of Northern China and Its Association with the Northern Hemisphere Circumglobal Teleconnection

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**ABSTRACT**

Using the latest daily observational rainfall datasets for the period 1961–2008, the present study investigates the interannual variability of June–September (JJAS) mean rainfall in northern China. The regional characteristics of JJAS mean rainfall are revealed by a rotated empirical orthogonal function (REOF) analysis. The analysis identifies three regions of large interannual variability of JJAS rainfall: North China (NC), Northeast China (NEC), and the Taklimakan Desert in Northwest China (TDNWC). Summer rainfall over NC is shown to have displayed a remarkable dry period from the late 1990s; while over NEC, decadal-scale variation with a significant decreasing trend in the last two decades is found, and over TDNWC, evidence of large interannual variability is revealed. Results also show that the interannual variability of JJAS rainfall in northern China is closely associated with the Northern Hemisphere circumglobal teleconnection (CGT). Correlation coefficients between the CGT index and regional-averaged JJAS mean rainfall over NC and NEC were calculated, revealing values of up to 0.50 and 0.53, respectively, both of which exceeded the 99% confidence level.

**Key words**: rotated empirical orthogonal function analysis, arid, semiarid, interannual variability, circumglobal teleconnection

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## 1. Introduction

The arid and semi-arid regions (ASAR) of China, mainly located in the north, cover nearly half of the land area (e.g. Chen et al., 2009). The annual rainfall in these regions is almost less than 400 mm. The amount of rainfall in most parts of ASAR, during the rainy months of June, July, August, September (JJAS), accounts for more than 70% of annual rainfall (Fig. 1). Rainfall is an essential component of the water balance. Its anomalies have crucial effects on industry and agriculture, as well as the lives of people, particularly for ASAR in China. Previous studies have shown signs of severe desertification and persistent drought in the semiarid region of northern China in recent decades (Wang and Li, 1990; Yatagai and Yasunari, 1994; Huang et al., 1999; Huang, 2006). The local and remote effects of the land surface processes and land degradation over ASAR in northern China have also been discussed in observational and numer-

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ical studies (e.g. Xue, 1996; Zhang et al., 2005; Zhou and Huang, 2006). These studies have indicated that variations in land-surface sensible flux and land degradation can cause local and remote rainfall anomalies. In turn, rainfall anomalies will affect land surface processes and cause changes to the environment. Therefore, identifying the long-term and interannual variability of rainfall and understanding the underlying mechanisms are of great importance to the future development of the economy, not only in ASAR, but also for other parts of the East Asian monsoon region. Much effort has been applied in studying summer rainfall variability in northern China during recent decades. Such work can generally be divided into two aspects, based on the domain of interest: persistent drought in North China, and the climate variations of ASAR in Northwest China.

For the first aspect, many previous studies have indicated there exists a decadal shift and a significant decreasing trend in summer precipitation over North China since the mid-1970s (Wang and Li, 1990; Chen et al., 1992; Yatagai and Yasunari, 1994; Yatagai and Yasunari, 1995; Nitta and Hu, 1996; Huang et al., 1999; Li et al., 2002; Gong et al., 2004; Xu et al., 2005; Ding et al., 2008; Huang et al., 2008). Wang and Li (1990) documented a possible link between precipitation fluctuations over the semiarid region of northern China and the summer monsoon, the western Pacific subtropical high, and ENSO. Huang et al. (2008) put forward a schematic model for how the interdecadal variations of the East Asian Climate system cause the persistent drought in North China, which is associated with ENSO, the subtropical High, mid-latitude wave train activity, Tibetan Plateau snow, and land surface processes. They proposed that the East Asian monsoon system was a coupling system of atmosphere, ocean, and land, and that the drought in North China (NC) was affected by the variability of East Asian monsoon climate subsystems, as well as their interaction between each other.

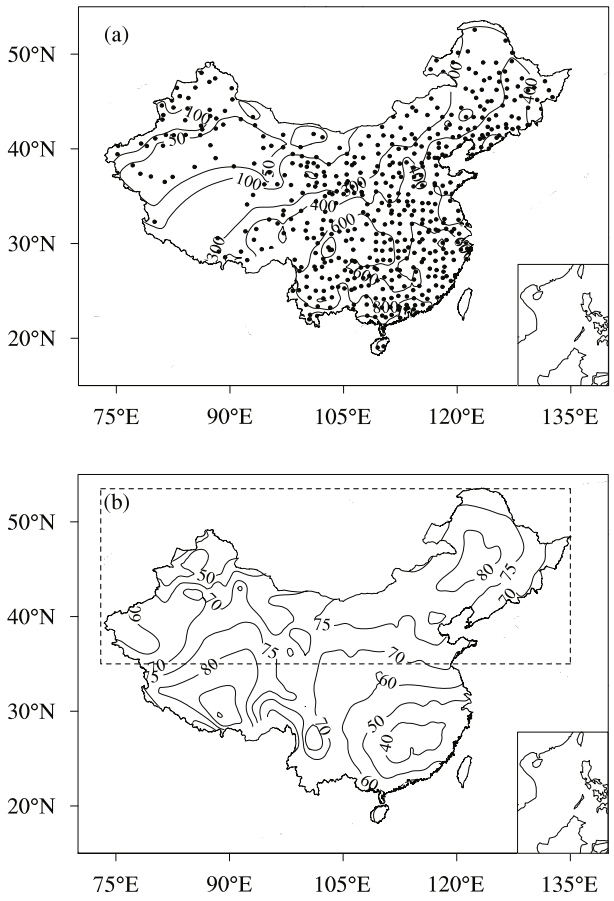



**Fig. 1.** (a) Geographical distribution of mean JJAS total precipitation (mm) and (b) the ratio (%) of JJAS mean rainfall to annual mean rainfall in China for the period 1961–2008. Dots in (a) denote the 552 weather stations. The dashed rectangular area represents the domain covered in the study.

For the second aspect, many observational studies have revealed that summer rainfall in Northwest China (NWC) shows an increasing trend since the 1960s, with a remarkable shift in the mid-1970s (Yatagai and Yasunari, 1994, 1995; Gong et al., 2004; Endo et al., 2005; Ma and Fu, 2006; Shi et al., 2007; Yang and Zhang, 2008; Zhou and Huang, 2009). These studies have also indicated that summer rainfall change in NWC exhibits regional characteristics, with a stage of relative increase in western parts of NWC, but a stage of drought in eastern parts, and that the intensity of the increasing trend of summer rainfall is greater in northern Xinjiang than in southern Xinjiang. Yang and Zhang (2008) analyzed the relationships between summer rainfall in Xinjiang and the Asian subtropical westerly jet stream (ASWJS). They revealed that the summer rainfall anomaly in Xinjiang had a notable correlation to the position of the ASWJS, as well as quasi-stationary wave activity. Zhou and Huang (2009) investigated the possible cause of interdecadal variability in summer rainfall in NWC and pointed out that an increase in near-surface temperature after 1978 enhanced convective instability, which in turn contributed to a strengthening of ascent and an increase in summer rainfall in NWC during the period 1978–2000.

ASAR has a large span in the east–west direction. The rainfall anomaly pattern in this region may mirror large-scale characteristics of atmospheric circulation and indicate internal connections between different parts. However, most previous work has focused on decadal-scale rainfall variability and on one part

of ASAR only. Few studies have addressed climatic variations over ASAR as a whole, which may provide essential information for understanding the nature of desertification in northern China. Therefore, one of the aims of the present paper is to investigate the interannual variability and regional characteristics of JJAS mean precipitation, and the associated atmospheric circulation, across the whole of ASAR.

The influence of atmospheric teleconnnection on boreal summer climate has been studied extensively. For example, a study by Nitta and Hu (1996) indicated that the Pacific–Japan (PJ) and Eurasia teleconnection patterns (EU) play very important roles in the variation of coupled patterns of summer rainfall and temperature in China, especially along the middle-lower reaches of the Yangtze River. More recent studies have suggested that the EU teleconnection is a bridge connecting the Indian summer monsoon (ISM) and the East Asian summer monsoon (EASM) (e.g. Wang et al., 2001; Wu and Wang, 2002). Lu et al. (2002) indentified a teleconnection pattern in July emerging from North Africa to East Asia along the westerly jet in the middle latitudes. They suggested that the teleconnection is a possible linkage of the EASM to the Indian monsoon, and even to subtropical heating anomalies over the Atlantic. Ding and Wang (2005) revealed a recurrent CGT pattern in the summertime mid-latitude circulation in the Northern Hemisphere. The CGT, actually a combination of the ISM–EASM teleconnection, the Silk-Road, and the Tokyo–Chicago express, is accompanied by significant rainfall and surface temperatures in the continental regions of Western Europe, European Russia, India, East Asia, and North America. ASAR is mainly located between 35°N and 55°N, which is on the path of the CGT. As revealed in this paper, the dominant mode of the JJAS mean rainfall in ASAR is closely associated with the CGT, which may suggest a new way to understand drought in northern China.

The remainder of the text is organized as follows. The datasets and methodology used are described in section 2. Section 3 presents the spatial and temporal variations of JJAS mean rainfall in northern China. In section 4, the relationship between the upper-level circumglobal teleconnection and JJAS rainfall in China is examined. A discussion is provided in section 5. And finally, conclusions are summarized in section 6.

## 2. Data and methods

### 2.1 *Datasets*

Two major datasets were used in this study. The first comprised daily observational rainfall data from 828 weather stations, and was provided by the National Meteorological Information Center of China, spanning the period 1951–2008. To achieve temporal continuity in terms of coverage from each station, we chose 552 stations (dots in Fig. 1a) from the complete data archive, as well as a slightly shorter period spanning 1961–2008 for this study. The second dataset comprised monthly atmospheric data for the period 1961–2008 from the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis (Kalnay et al., 1996), on a horizontal resolution of 2.5°×2.5°. The variables used in this study included winds and height at the 200, 500, 700, and 850 hPa levels.

### 2.2 *Methods*

As already mentioned, the present study concerns the area of northern China; however, more specifically, this covers the domain (35°–55°N, 73°–136°E), which includes a total of 273 weather stations (area denoted by the dashed rectangle in Fig. 1b). The rotated empirical orthogonal function (REOF) method of analysis was applied to reveal the regional characteristics of interannual variations in JJAS mean precipitation over the domain concerned. Firstly, the station datasets were normalized and long-term trends removed. Then, a correlation matrix of the rainfall data was used to compute the eigenvalues and eigenvectors of EOFs. After that, the first three EOFs were rotated orthogonally from the six leading modes of EOFs. Regression and correlation analysis were also used, in order to study the relationship between the interannual variation of JJAS rainfall in northern China and general atmospheric circulations.

## 3. Spatial and temporal variations of JJAS mean rainfall in northern China

### 3.1 *Long-term trend of JJAS mean rainfall in northern China* (1961–2008)

This study employed a relative variation ratio (RVR) to indicate the extent of change in JJAS mean rainfall during the period 1961–2008 in northern China. The RVR was obtained by dividing the linear trend of JJAS mean rainfall of one station by the JJAS mean rainfall of that station. The distribution of RVR in China is displayed in Fig. 2. A notable west–east contrast, with rainfall increasing in most parts of NWC, but decreasing in NC, Northeast China (NEC), can be observed. Furthermore, the largest positive and negative centers exist in North Xinjiang, and NC, respectively. These results are consistent with previous studies.

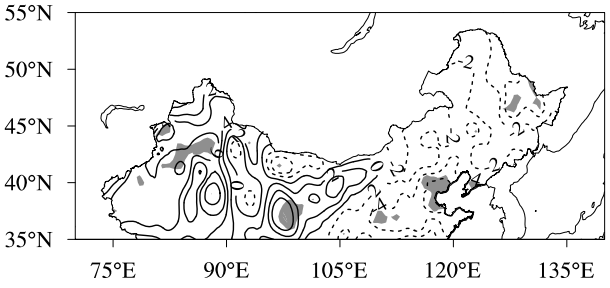



**Fig. 2.** RVR (%) of JJAS mean rainfall in northern China over the period 1961–2008 [units: mm (10 yr)$^{-1}$]. Shading indicates regions with trend anomalies beyond the 90% confidence level.

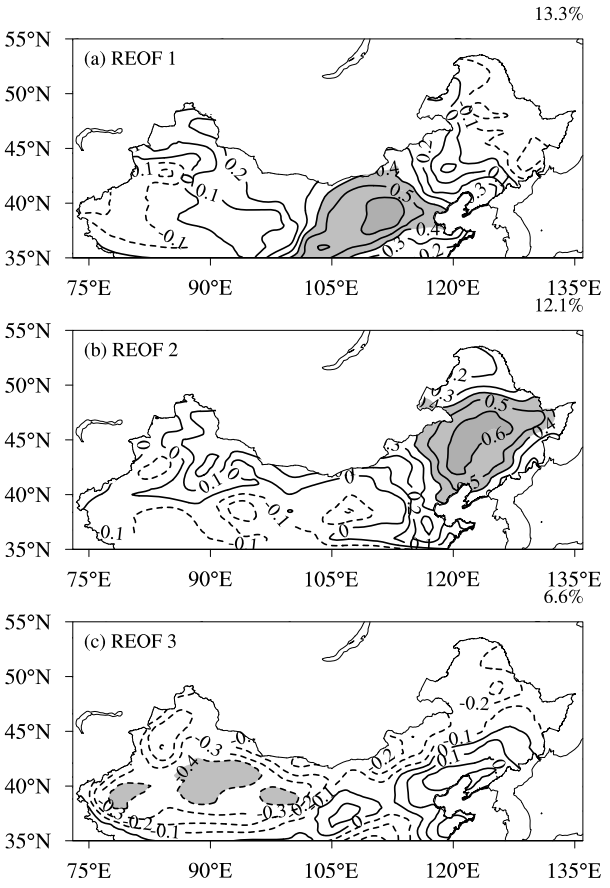



**Fig. 3.** Spatial modes of the first three REOFs of JJAS mean rainfall anomalies in northern China (contour interval: 0.1). Shaded areas indicate regions with values beyond 0.4 and less than −0.4. The contribution of each mode to the total variance is displayed in the upper right of each panel.

### 3.2 *Spatial and temporal variation of JJAS mean rainfall in northern China* (1961–2008)

To investigate the interannual variation in JJAS mean precipitation in northern China, REOF analysis, a widely used diagnostic tool in climatology, was applied. It retains the advantages but avoids the spatial limitations of EOF analysis, and can reveal efficiently the regionality of climatic factors (e.g. Richman, 1986; Murata, 1990; Yatagai and Yasunari, 1995; Li et al., 2002). The spatial patterns of the first three leading REOFs of the JJAS mean rainfall in northern China are shown in Fig. 3. The three modes account for 13.3%, 12.1%, and 6.6% of the total variance, respectively. The first two REOF modes feature a quasi-zonal tripole-like distribution with an anomalous center over the horseshoe-shaped arc region and adjacent region in between. The pattern in Xinjiang Province shows an opposing north–south distribution, which agrees with Yang and Zhang (2008). The third mode displays a consistent anomalous pattern in most

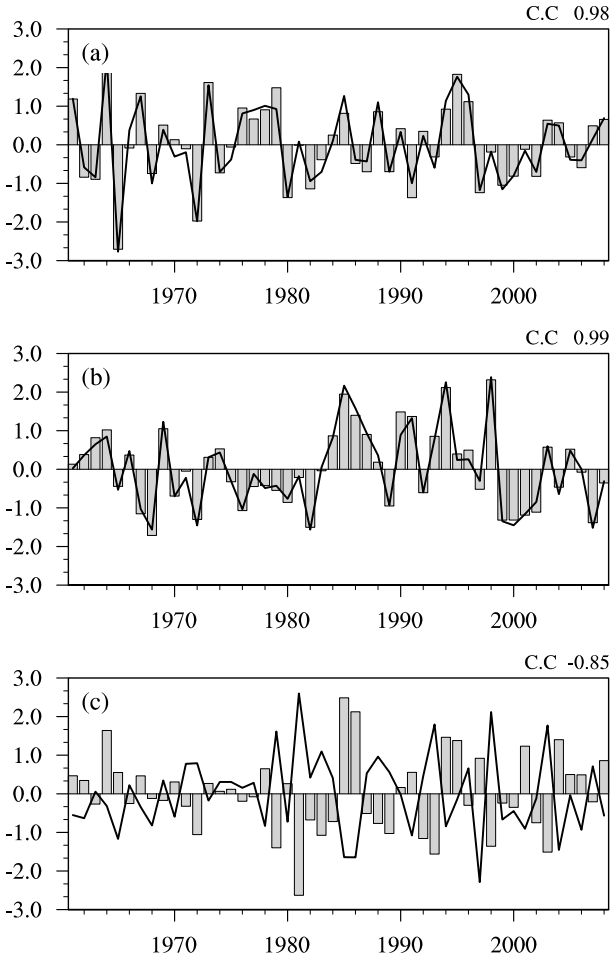



**Fig. 4.** Normalized time coefficients of each REOF mode [histograms: (a) for REOF1, (b) REOF2, and (c) REOF3] and the normalized series of regional mean JJAS rainfall for each relevant region [solid lines: (a) for NC, (b) NEC, and (c) TDNWC]. And the correlation coefficient (C.C) between the two time series in each subpanel is also added in its upper right.

parts of northern China, especially in NWC.

The spatial distribution in Fig. 3 designates the correlation coefficients of JJAS mean rainfall anomalies at each station with the time series corresponding to the REOF modes. The shaded areas, with values beyond 0.4 or less than −0.4, denote the correlation coefficients exceeding the 99% confidence level. So, three regions (shaded in Fig. 3) associated with the three dominant modes can be deduced: most parts of NC, NEC, and the Taklimakan Desert region in northwest China (TDNWC), with 76, 67, and 28 weather stations in each region, respectively.

The corresponding time coefficient of each principle component (PC) to the three REOF modes (histogram) and the JJAS mean rainfall anomalies over the regions of NC, NEC, and TDNWC (solid lines) are depicted in Fig. 4. It is obvious that the three PCs match well with the regional averaged JJAS mean rainfall anomaly of the associated region. The correlation coefficients between the three pairs of time series are 0.99, 0.98, and −0.86, respectively, all beyond the 99% confidence level based on Pearson's $r$ statistical significance test. Thus, the three PCs characterize JJAS mean rainfall variability of the corresponding region. Furthermore, by combing the power spectrum of each PC (Fig. 5), each of their features of variability can be obtained. PC1 exhibits a 3- and 9-year periodic variation, showing a remarkable relatively dry period from the mid-1990s to the present day. PC2 shows decadal-scale variation and a distinct decreasing trend in the last two decades, with dry periods from the late 1960s to the early 1980s, and from the late 1990s to the present day, and with wet periods between. Meanwhile, there is no significant frequency based on the spectral analysis, as shown by Fig. 5b. For the third PC, the spectral analysis result reveals large interannual variability, with 4–5- and 9-year periods of variation (Fig. 5c).

## 4. The relationship between the CGT and interannual variations of JJAS mean rainfall in northern China

The CGT is a geographically phase-locked recurrent teleconnection pattern during boreal summer from July to September, having its path within the westerly jet stream from the Northeast Atlantic through Eurasia, the North Pacific, to North America, and having three center cells primarily in East Asia, mainly in northern China (Ding and Wang, 2005). Interestingly, the interannual variability of JJAS mean rainfall in northern China shows a clear quasi-zonal wave-like distribution in the first two REOFs, with the center over a horseshoe-shaped arc region, where the transition area of ASAR is located. It is conceivable that the first two leading REOFs may have a potential connection with the CGT. This section investigates this relationship.

### 4.1 *The CGT patterns associated with the first two leading modes of JJAS mean rainfall in northern China*

To begin with, the atmospheric circulation anomalies corresponding to the first two dominant modes of the REOFs are explored using the regression method. Figure 6 shows the regressed patterns of JJAS mean height at the 200 hPa and 700 hPa levels with respect to the first two time coefficients mentioned above. The

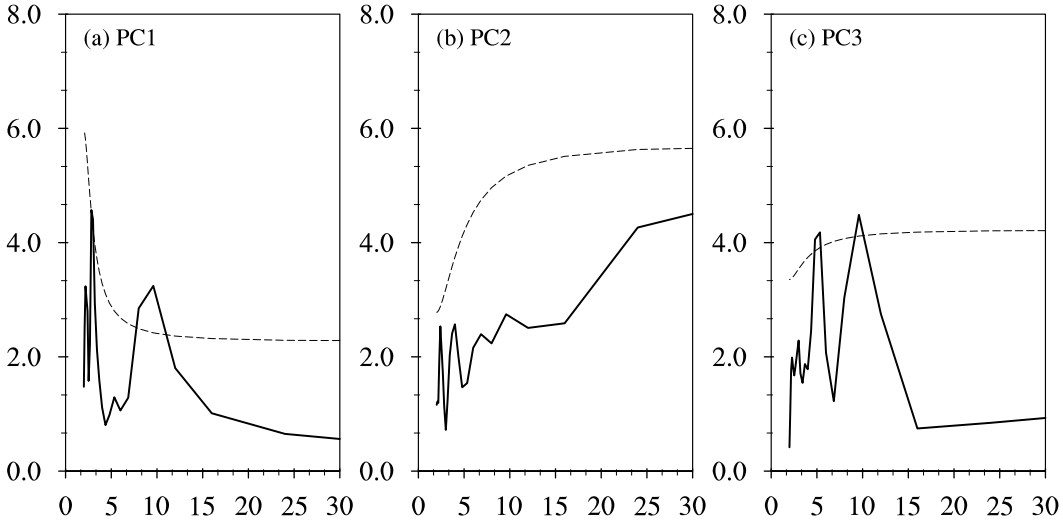



**Fig. 5.** The power spectrum (solid line) for the principle component (PC) of the three REOFs, and the dotted thin lines represent the power spectrum at the 95% confidence level.

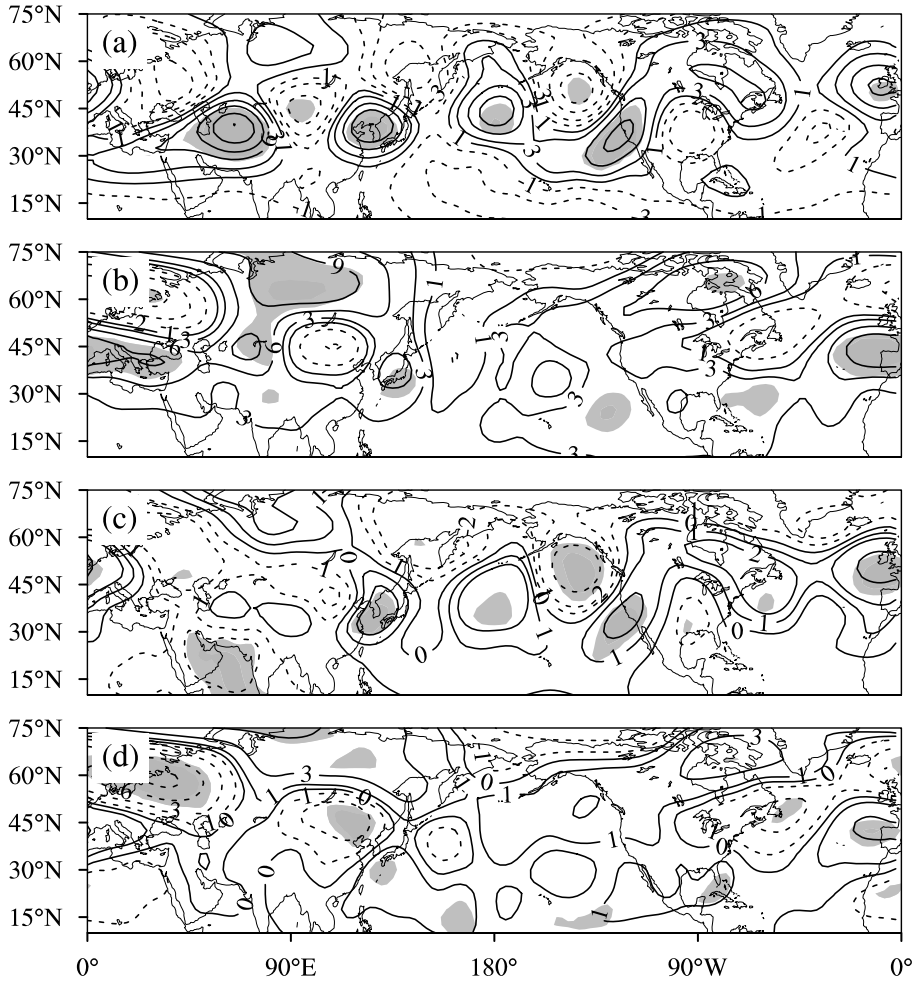



**Fig. 6.** Anomalies of regressed geopotential height at 200 hPa (a, b) and 700 hPa (c, d). (a, c) against PC1; (b, d) against PC2. Shading represents regions with height anomalies beyond the 95% confidence level.

most prominent feature in Fig. 6a, which is related to PC1 regressed onto the height at 200 hPa, is a wave train structure from the Northeast Atlantic through Eurasia, the North Pacific, to North America, with positive anomalous centers over the Northeast Atlantic, central Asia, Northeast Asia, the central-north Pacific, and western North America, with negative centers between. The largest positive center is found over central Asia in the domain (35°–40°N, 60°–70°E).

The regressed height pattern at the 200 hPa level against PC2 is shown in Fig. 6b. A similar wave train structure emerges, although not as clear and strong as that in Fig. 6a, especially over the North Pacific and North America. The positions of the anomalous centers move a little eastward compared to those in Fig. 6a. In addition, a NAO-like (North Atlantic Oscillation) pattern presents along the coastline of Western Europe from North Africa to the North Atlantic, suggesting influence from the high latitudes.

The atmospheric circulation at the 500 hPa level related to the two REOF modes exhibits a good similarity to that at the 200 hPa level (Figs. 6a, b), having anomalous centers over the same locations, but the amplitude of the anomalies increases with height (figures not shown).

The regressed geopotential height anomalies at 700 hPa corresponding to the two REOFs modes are shown in Figs. 6c and d. The regressed low-level height anomalies exhibit wave train patterns similar to those shown in Figs. 6a and b, though with weaker intensity and a tilted baroclinic structure over the northwest part of India where there is an anomalous ridge at 200 hPa and an enhanced cyclonic anomaly over the Arabian Peninsula and adjacent sea at 700 hPa. The regressed wind fields against the two REOF modes match the height anomalies well (figure not shown).

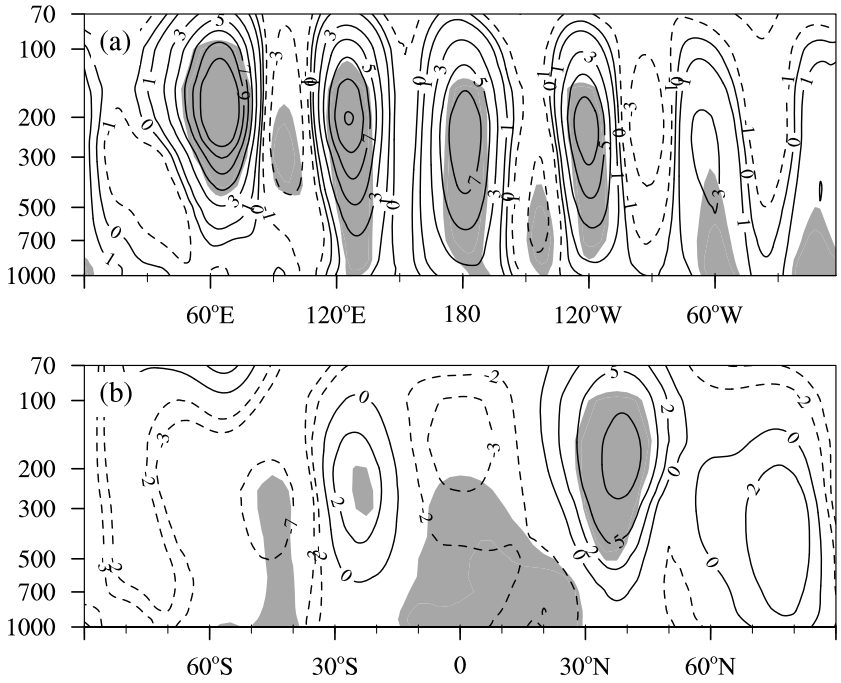



**Fig. 7.** (a) Height–longitude cross-section anomalies of PC1 regressed onto height averaged within 30°–50°N. (b) Height–latitude cross-section anomalies of PC1 regressed onto height averaged within 60°–70°E. Shaded reveals the regions with height anomalies exceed the 90% confidence level.

The vertical structure of the wave train pattern can also be examined. Figure 7a illustrates the height–longitude cross section anomalies of PC1 regressed onto height averaged within 30°–50°N. Evidently, a wave train structure emerges along the westerly jet from near 60°E, passing over Eurasia to North America. This wave-like pattern shows a vertically quasi-barotropic structure for nearly all the wave train centers, except for a tilted baroclinic pattern over the northwest part of India. Also, it has a wavenumber-6 structure, with an intense region of positive/negative anomalous centers confined to 30°–50°N within a waveguide associated with the westerly jet stream extending from Eurasia to North America. Furthermore, it is evident that the maximum anomalous center for each cell is at about the 200 hPa level, with the largest one occurring around 65°E. These features demonstrate a geographically fixed global-scale teleconnection pattern, strikingly similar to the CGT revealed in Ding and Wang (2005). Moreover, the regressed patterns at 200 hPa are identical to the two scenarios of the CGT proposed by Ding and Wang (2005) [see Fig.15 in Ding and Wang (2005)], respectively. These similarities can be clearly seen in Fig. 8.

Figure 8 displays the meridional wind anomalies associated with the first two PCs. The pattern corresponding to PC1, as shown in Fig. 8a, exhibits an apparent wave-like structure propagating within the westerly jet stream from the Northeast Atlantic, through Eurasia and the North Pacific, to eastern America, and for the regressed pattern against PC2 shown in Fig. 8b, a notable teleconnection pattern emerges over the Eurasian continent from Northern Europe to East Asia. Additionally, a distinct discrepancy is observed over the Arctic region in which the magnitude of anomalies regressed against PC1 are much weaker than those against PC2. This implies that the strong anomalous activity over the Arctic area, or from high latitudes, may affect the wave-like anomaly pattern. The features revealed in Fig. 8 matches well with Ding and Wang (2005).

### 4.2 *JJAS rainfall anomalies in China associated with CGT patterns*

The results thus far have indicated that the interannual variability of JJAS mean rainfall in northern China shows a considerable correlation with the CGT pattern. Here, following Ding and Wang (2005), the correlation distributions between them are examined. First, two indices, CGTI-1 and CGTI-2, are defined to represent the two scenarios of CGT variability, respectively.

For the first scenario, we employ the definition by Ding and Wang (2005), which is defined as the norma-

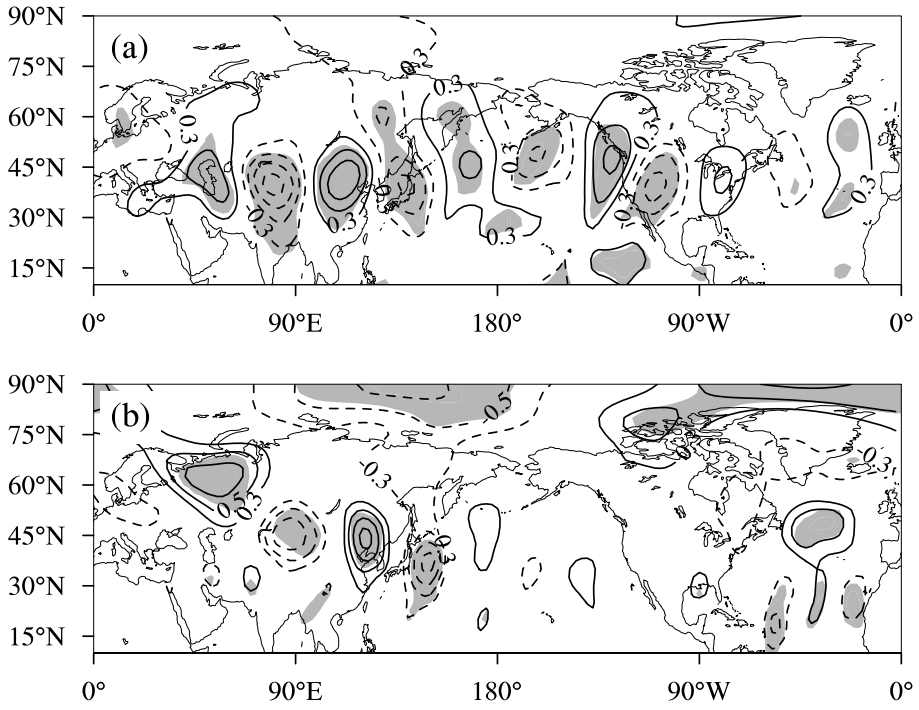



**Fig. 8.** Anomalies of JJAS mean meridional wind at 200 hPa regressed against PC1 (a) and PC2 (b). Shallow area indicates the region with wind anomalies beyond the 95% confidence level.

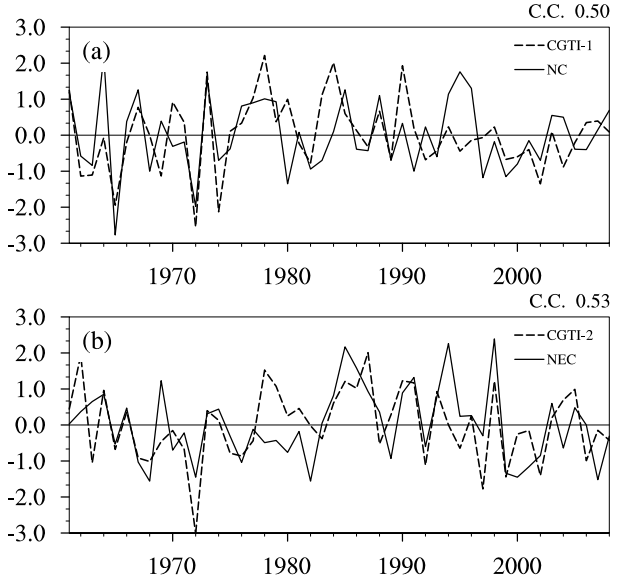



**Fig. 9.** Normalized time series of CGTI 1-2 (dashed lines) and regional mean JJAS rainfall in NC and NEC (solid lines). And the correlation coefficient (C.C) between each pair of time series is also added in the upper right of each panel.

zlied time series of the interannual variability of the 200 hPa height averaged over the reference area (35°–40°N, 60°–70°E), and denote it as CGTI-1. CGTI-2, corresponding to the second scenario, is defined using the difference between the normalized time series of regionally-averaged height at 200 hPa over the two areas, (40°–50°N, 20°–10°W) and (60°–65°N, 15°–25°E). The regressed height anomalies at 200 hPa against the two indices resemble those in Fig. 6 (figures not shown).

Figure 9 displays the two indices and the normalized time series of regionally-averaged JJAS mean rainfall in NC and NEC. The correlation coefficients between the two pairs of time series are 0.50 and 0.53, respectively, both beyond the 99% confidence level. The correlation coefficients of the JJAS mean rainfall and the two CGTIs are displayed in Fig. 10, the patterns of which resemble the spatial modes of the first two REOFs (Figs. 3a, b), with significant anomalies located in northern China, especially NC, NEC, and western part of Xinjiang Province. The composite atmospheric circulations related to the CGTIs are also performed. Figure 11 shows the positive-minus-negative CGTI composite wind anomalies at 850 hPa (vector) and vertical velocities at 500 hPa (contour), the cases of which are chosen from the lower and upper bounds, respectively, with a 0.5 standard deviation of the CGTIs. The composite results show that, corresponding to each scenario, an anomalous southerly

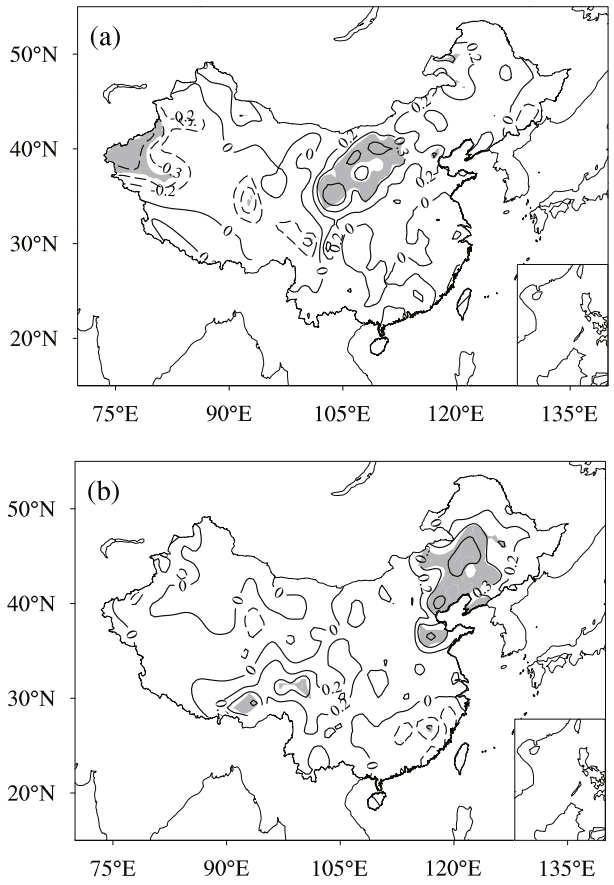



**Fig. 10.** Correlation coefficients between CGTI and the JJAS mean rainfall in China: (a) CGTI-1; (b) CGTI-2. Shaded area illustrates the region of rainfall anomalies beyond the 95% confidence level.

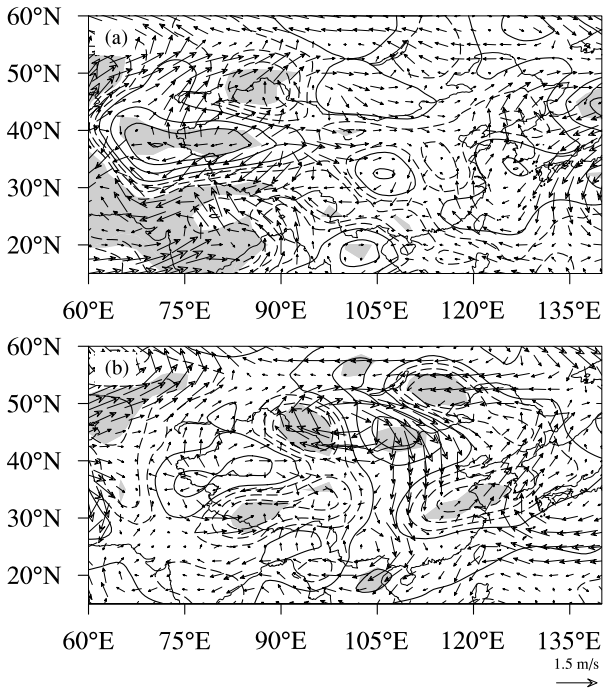



**Fig. 11.** Composite anomalies of winds (vectors, arrow with scale at bottom right of the panels) at 850 hPa and vertical velocity (contour) at 500 hPa associated with the two CGTIs: (a) CGTI-1; (b) CGTI-2. Shading denotes the region with vertical velocity anomalies beyond the 95% confidence level.

over Eastern China with abundant moisture flows to northern China converges with the cold air from high latitudes and triggers anomalous ascending motion over NC/NEC, all of which are most favorable for the occurrence of rainfall, and thus positive rainfall anomalies are found over the two regions. Meanwhile, both the two CGTIs and the rainfall anomalies in NC/NEC are in a below-normal phase during recent decades. As to why this happens requires further study.

## 5. Discussion

Previous studies have already revealed that there are teleconnection patterns and stationary Rossby waves propagating eastward within the Asian jet in boreal summer (e.g. Ambrizzi et al., 1995; Rodwell and Hoskins, 1996; Krishnan and Sugi, 2001; Lu et al., 2002; Wu and Wang, 2002; Enomoto et al., 2003; Enomoto, 2004; Ding and Wang, 2005; Sato and Takahashi, 2006) and these features can be achieved by a few variables, such as geopotential height, streamfuction, and meridional wind. The formation mechanism of the wave-like pattern on the jet has also been discussed (Rodwell and Hoskins, 1996; Enomoto et al., 2003; Ding and Wang, 2005; Sato and Takahashi, 2006; Shi and Lu, 2010). Rodwell and Hoskins (1996) revealed that the diabatic heating associated with the Indian monsoon resulted in diabatic cooling and downward motion over the Eurasian continent and Mediterranean Sea. A study by Enomoto et al. (2003) indicated that heating from the Indian monsoon and the related diabatic forcing over the continent would excite a quasi-stationary disturbance on the jet, triggering a barotropic structure propagating within the westerly jet. Sato and Takahashi (2006) reported that a Rossby wave packet propagated eastward from the Middle East, passing over East Asia, and reaching North America. They showed that the anomaly pattern was strengthened through kinetic energy conversion near the entrance of the Asian jet over the Middle East, and that the interaction between the anomaly pattern and the basic field contributed to the appearance of the anomalous wave-like pattern.

Most of the studies mentioned above discuss the teleconnection pattern on the monthly or submonthly

timescale. On the seasonal scale, the features of the teleonnection also, to some extent, reveal the possible relationship between the teleconnection and associated factors mentioned above, such as the westerly jet (shown in Fig. 7a) and the heating from the ISM. Figure 7a shows a clear wave train pattern propogating eastward along the westerly jet, which plays the role of the waveguide confining the wave train to within the domain 30°–50°N. Furthermore, interestingly, the teleconnection pattern exhibits an evident baroclinic structure around 65°E, with a maximum center at about 200 hPa. Thus the vertical structure around 65°E is investigated. Figure 7b depicts the height–latitude cross section of geopotential height anomalies (averaged within the domain 50°–70°E) regressed against PC1. A barotropic teleconnecton pattern appears with out-of-phase anomalies between the tropics and extratropics, especially in the Northern Hemisphere. This structure can also be seen in Fig. 11a, where the vertical velocity anomalies at 500 hPa show an "ascent–descent–ascent" pattern around 65°E. Furthermore, we can determine that the meridional teleconnection pattern may be closely related to the low level activity over the Indian Continent/Ocean, where significant height anomalies are apparent. This meridional teleconnection pattern resembles the low-frequency waveguide (or a biweekly oscillation) at 200 hPa near 60°E, mentioned in Liu and Yu (1993). These authors documented that the wave source of this waveguide was located around 10°–25°N, and the perturbation propagated polarward from the wave source to affect the westerlies of the Southern and Northern Hemispheres. This indicates that the waveguide near 60°E may be a bridge connecting the tropics and extratropics. Wang et al. (2005) proposed a mechanism for how the tropical activity affected the extratropical atmosphere. They indicated that a forcing embedded in the deep tropical easterlies may excite a Rossby wave response in the extratropical westerlies and the southerly flow component in the basic state played a role of a conveyor, transferring a Rossby wave source northward. The study by Wang et al. (2005) also showed that the southerly conveyor determined the location of the effective Rossby wave source and that the extratopical response is relatively insensitive to the location of the tropical forcing, provided that the tropical response can reach the southerly conveyor. A stronger southerly flow favors a stronger extratropical response, and the spatial structure of the extratropical response is determined by the extratropical westerly basic flows. This result agrees well with Sato and Takahashi (2006), that the appearance of the Rossby wave on the westerly jet is clearly phase-dependent, and may be dominanted by the internal dynamics of interaction between the anomaly pattern and the basic flows. This implies that the anomalous convectivity over the Indian Ocean, or diabatic heating from the ISM, exicites a Rossby wave response in the extratropical westerly by triggering the anomalous southerly in the tropics, thus affecting the wavelike pattern on the jet. However, these results were achieved based on a simple barotropical model or statistical analysis, and the real atmosphere is much more complicated. The impact of tropical acitivity on the wavelike pattern on the jet is still affirmative.

## 6. Conclusions

Using the latest daily observation rainfall data sets for the period 1961–2008, the trend and interannual variations of JJAS mean rainfall in northern China have been studied. Atmospheric circulation anomalies associated with the interannual variation of summer rainfall have also been explored. The major findings are summarized as follows.

Firstly, the RVR of the JJAS rainfall in northern China over the period 1961–2008 shows a "west–increase–east–decrease" pattern, with considerable anomalies over Xinjiang Province and North China. Secondly, the spatial and temporal structures of the interannual variation of JJAS mean rainfall in northern China were revealed by REOF analysis. The first two dominant modes showed a quasi-zonal tripole-like pattern, with the centers over the transition areas of ASAR in northern China. Three regional divisions were identified by the REOF analysis: NC, NEC, and TDNWC. Results indicated that the variation frequency of the summer rainfall anomaly varies between regions. Summer rainfall over NC revealed two dominant variation frequencies and a remarkable dry period from the late 1990s. Over NEC, decadal-scale variation and a decreasing trend in the last two decades was found, but with no significant frequency variation. TDNWC exhibited large interannual variability.

This study has also demonstrated that interannual variations of JJAS mean rainfall in northern China are well related to the upper-level CGT in boreal summer, especially in NC and NEC. The significant summer rainfall anomalies related to the CGT were found to be closely connected with the locations of the action centers of the CGT. There are three center cells of the CGT primarily over East Asia, mainly in northern China, which affect moisture transportation and ascending motion over these regions. Results showed that a positive CGTI will excite a southerly anomaly, bringing more moisture to northern China; meanwhile, upper-level divergence favors low-level convergence and ascending motion over

NC/NEC. As such, positive rainfall anomalies tend to be found in NC/NEC. The correlation coefficients of the regional averaged JJAS mean rainfall over NC and NEC with the CGTI were up to 0.50 and 0.53, respectively, both beyond the 99% confidence level.

The CGT pattern reveals a geographically phase-locked structure during boreal summer from July to September (Ding and Wang, 2005) and is accompanied by significant climate anomalies along its path. This is confirmed by the present study. The closely in-phase relationship between the rainfall anomalies in NC/NEC and the teleconnection patterns provides us with a new way to understand the persistent drought in NC and the decreasing rainfall in NEC. Finally, what contributes to the long-term negative phase of CGTI, and whether the wavelike pattern is the dominant factor contributing to the long-term decrease in northern China, which remain unclear and require further investigation.

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# An Index Measuring the Interannual Variation of the East Asian Summer Monsoon—The EAP Index

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**ABSTRACT**

Based on the EAP (East Asia/Pacific) teleconnection in the summer circulation anomalies over the Northern Hemisphere, an index measuring the strength of the East Asian summer monsoon, i.e., the so-called EAP index, is defined in this paper. From the analyses of observed data, it is clearly shown that the EAP index defined in this study can well describe the interannual variability of summer rainfall and surface air temperature in East Asia, especially in the Yangtze River valley and the Huaihe River valley, Korea, and Japan. Moreover, this index can also reflect the interannual variability of the East Asian summer monsoon system including the monsoon horizontal circulation and the vertical-meridional circulation cell over East Asia. From the composite analyses of climate and monsoon circulation anomalies for high EAP index and for low EAP index, respectively, it is well demonstrated that the EAP index proposed in this study can well measure the strength of the East Asian summer monsoon.

**Key words:** East Asia/Pacific index, East Asian summer monsoon, interannual variability

## 1. Introduction

As well known, the interannual variability of the East Asian summer monsoon is very obvious and causes severe droughts and floods in East Asia, especially in the eastern part of China (see Tao and Chen, 1985; Ding, 1994; Ye et al., 1996; Huang et al., 1998; Huang and Zhou, 2002). Therefore, the interannual variability of the East Asian summer monsoon (EASM) is an important issue for Chinese meteorologists. Through the efforts of more than sixty years, Chinese meteorologists have achieved many advances in the studies on the EASM. Tao and Chen (1987) made a systematical review on these advances.

Recently, many scholars have made many efforts to study the interannual variability of EASM and its causes. Nitta (1987, 1996), Huang and Li (1987, 1988), Kurihara (1989), and Huang and Sun (1992) pointed out the significant effect of the thermal state in the tropical western Pacific and the convective activities around the Philippines on the interannual variability of EASM. Nitta (1987), and Huang and Li (1987) showed that there is an obvious teleconnection pattern in the summer circulation anomalies over the Northern Hemisphere, which has been called the PJ (Pacific-Japan) pattern and the EAP (East Asia/Pacific) pattern teleconnection by Nitta (1987), and Huang and Li (1987), respectively.

In order to describe the interannual variability of EASM, a monsoon index measuring the strength of EASM is necessary. Up to now, there are two kinds of definitions of the Asian monsoon index. One is defined from the thermodynamical elements. For example, Tao and Chen (1987), and Murakami and Matsumoto (1994) defined the monsoon index with the strength of monsoon rainfall and OLR, respectively. The other is defined from the dynamical elements, i.e., wind field. Webster and Yang (1992), and Zeng et al. (1994) defined the monsoon index from the difference of the zonal winds between the high level and lower level atmosphere over the Asian monsoon region, while Zhang and Peng (2003) defined the East Asian summer monsoon index with the difference of the zonal winds between the middle latitudes and low latitudes over East Asia. There are some advantages and disadvantages in these definitions. The former may be easily influenced by local thermodynamical conditions, but the latter may be only suitable for the South Asian monsoon. Since the EASM has the feature of a subtropical monsoon and a meridional component (see Huang et

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al., 1998), it may be not suitable to define an index only with the zonal component of the wind field over the regions of EASM. Therefore, how to define a monsoon index measuring the strength of the EASM is still necessary for the study of the interannual variations of EASM.

Guo (1983) and Shi et al. (1996) defined the index measuring the interannual variations of EASM with the difference of sea-level pressure between the North Pacific and Eurasian continent, which is also an element related with the dynamics. As described above, the observational and theoretical studies have shown that the interannual variation of EASM is greatly influenced by the EAP teleconnection pattern suggested by Nitta (1987) and Huang and Li (1987), respectively. This teleconnection pattern has been widely used in the seasonal prediction of climate anomalies in China. Thus, based on this teleconnection shown in Figure 1 (from Huang and Li, 1987), the NCEP/NCAR reanalysis data of height fields at 500 hPa are used to defined an index measuring the strength of the EASM. And the observed data of monthly precipitation in China, Korea, and Japan are further used to study the interannual variations of EASM in this paper.

## 2. Definition of the EAP index measuring the interannual variation of EASM.

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of

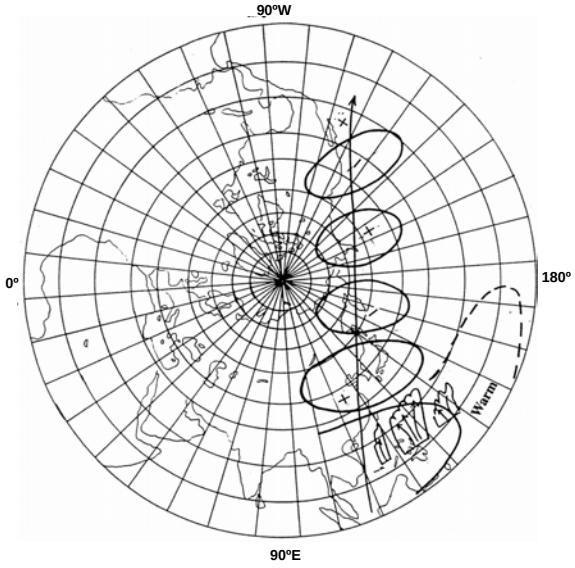



**Fig. 1.** The schematic map of the EAP teleconnection pattern (from Huang and Li, 1987). The "+" and "–" in the figure denote the positive and negative height anomalies at 500 hPa, respectively.

Fig.1

EASM, an index of EASM is defined based on this pattern. This index is simply called the EAP index, i.e., $I_{\rm EAP}$. The definition of $I_{\rm EAP}$ is as follows:

$$I_{\rm EAP} = -0.25Z'_{\rm s}(60^\circ{\rm N}, 125^\circ{\rm E}) + 0.50Z'_{\rm s}(40^\circ{\rm N}, 125^\circ{\rm E}) - 0.25Z'_{\rm s}(20^\circ{\rm N}, 125^\circ{\rm E}) \,, \quad (1)$$

where $Z'_{\rm s}{=}Z\sin 45^\circ / \sin\varphi$ is the standardized seasonal-mean 500-hPa height anomaly at a grid point with the latitude $\varphi$, and $Z'{=}Z-\bar{Z}$ is the summer seasonal-mean 500-hPa height anomaly at the grid point.

Figure 2 shows the interannual variation of $I_{\rm EAP}$ calculated using Formula (1) with the NCAR/NCEP reanalysis data of the 500-hPa height fields from 1951 to 2000.

The results of Nitta (1987), Huang and Li (1987), Kurihara (1989), and Huang and Sun (1992), show that the convective activities are strong around the Philippines in the warming state of the West Pacific warm pool; in this case, the western Pacific subtropical high shifts northward, and the summer monsoon rainfall may be below normal in the Yangtze River and Huaihe River valley of China, and in South Korea and Japan. Then the $Z'_{\rm s}$ at the grid point of (40°N, 125°E) is positive, and $Z'_{\rm s}$ at the grid point of (60°N, 125°E) and (20°N, 125°E) are both negative. Thus, $I_{\rm EAP}$ will be positive. Therefore, if the EAP index shown in Fig. 2 is positive, the East Asian summer monsoon rainfall may be weak in the Yangtze River and Huaihe River valley of China, and in South Korea and Japan. On the other hand, in the cooling state of the West Pacific warm pool, the convective activities are weak around the Philippines. Then the western Pacific subtropical high shifts southward, and the summer monsoon rainfall may be above normal in the Yangtze River and Huaihe River valley of China, and in South Korea and Japan. In this case, the $I_{\rm EAP}$ shown in Fig. 2 is negative. Therefore, if the index is negative, the East Asian summer monsoon rainfall may be strong in the Yangtze River and Huaihe River valley of China, and in South Korea and Japan.

## 3. Relationship between the EAP index and the summer monsoon rainfall in East Asia

Ramage (1971) pointed out that monsoon exhibits not only wind reversal, but also the seasonal precipitation criterion. Therefore, a good monsoon index should indicate well the variation of precipitation in the monsoon region. In order to investigate whether the EAP index can well describe the interannual variations of EASM or not, the point correlations between the summer monsoon rainfall anomalies in China and

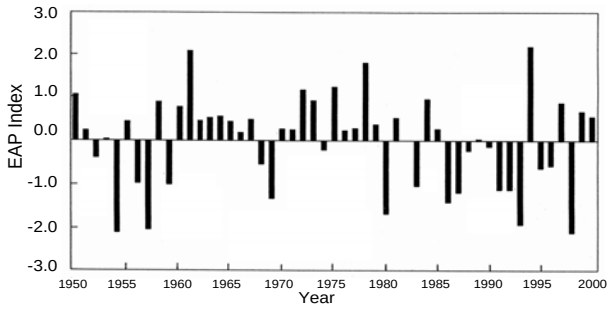



**Fig. 2.** Interannual variation of $I_{\mathrm{EAP}}$ calculated by using Formula (1) with the observed data of the 500-hPa height fields from 1951 to 2000.

the EAP index are calculated by using the observed data of summer (June–August) monthly precipitation at 160 observational stations in China from 1951 to 1996. Figure 3 presents the distribution of the correlation coefficients between the summer rainfall anomalies in China and the EAP index. It may be clearly seen from the figure that there are large negative correlations in the Yangtze River valley and the Huaihe River valley. This means that in a summer with positive EAP index the summer rainfall is below normal. For example, in the summers of 1961, 1978, and 1994, the EAP index was larger than or equal to 1, and correspondingly, severe droughts occurred in this region. However, in a summer with negative EAP index such as in 1954, 1957, 1980, 1987, 1996, and 1998, the EAP index was smaller than or equal to –1, and serious floods were caused in this region.

This negative correlation between the EAP index and summer rainfall anomalies in the Yangtze River valley and the Huaihe River valley is also seen from the interannual variations of the normalized summer rainfall anomalies in the (29°–33°N, 114°–125°E) region of the Yangtze River and Huaihe River valley and the EAP index shown in Fig. 4. It shows that there is a good out of phase relationship between the summer rainfall in the Yangtze River and Huaihe River valley and the EAP index. The maximum correlation coefficience between them can reach 0.58, which greatly exceeds the confidence level of 95%. This indicates that the EAP index can well reflect the summer monsoon rainfall anomalies in the Yangtze River valley and the Huaihe River valley.

The EAP index can reflect the interannual variation of summer monsoon rainfall in South Korea. Figure 5 shows the normalized summer rainfall anomalies in South Korea and indicates a negative correlation between the summer rainfall anomaly in South Korea and the EAP index. The maximum negative correlation coefficient can reach –0.37, which is significant at the confidence level of 95%.

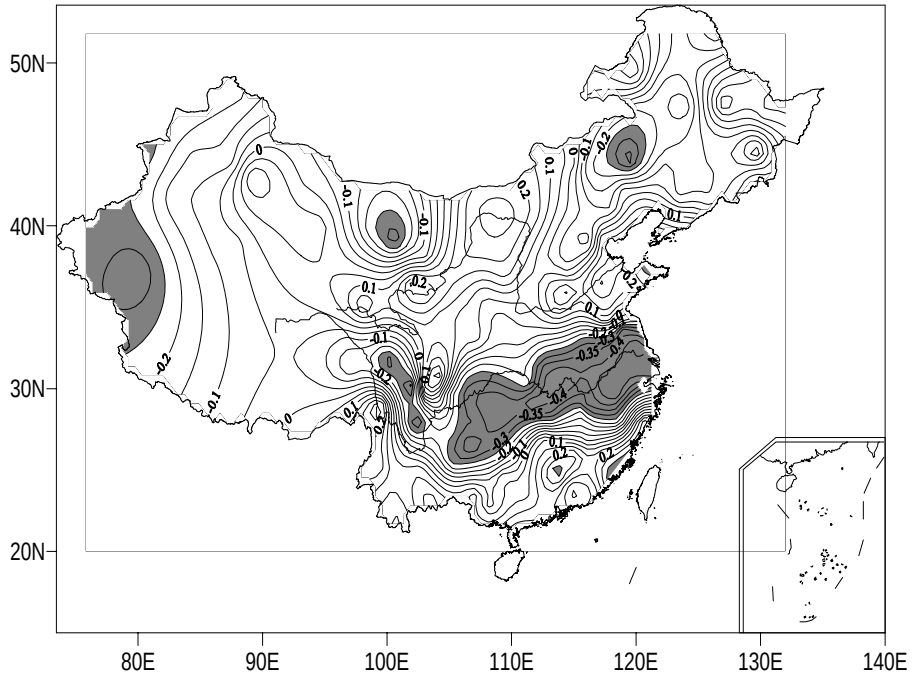



**Fig. 3.** Distribution of the correlation coefficients between the summer (June–August) rainfall anomalies in China and the EAP index from 1951 to 1996. The shaded areas indicate where the confidence level is over 95%.

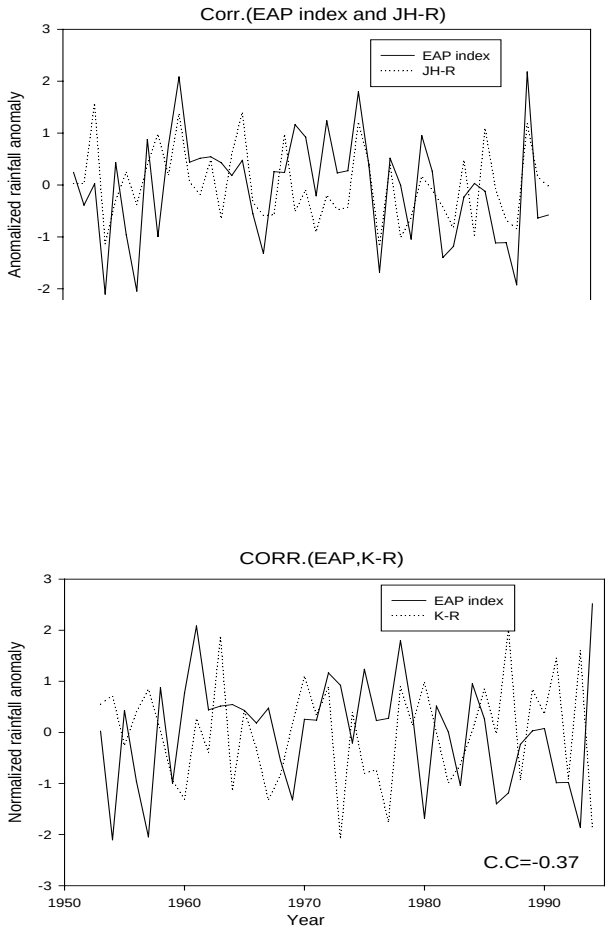



**Fig. 5.** Same as Fig. 4 but for the normalized summer monsoon rainfall anomalies in South Korea (dashed line).

We also calculate the correlation coefficients between the summer monsoon rainfall in East Asia and the EAP index, the summer monsoon index (SM index) suggested by Guo (1983), and the Webster-Yang index (WY index) proposed by Webster and Yang (1992), respectively, by using the Xie-Arkin precipitation data for 1979 to 1998.

Figures 6a–c are the distributions of the correlation coefficients between the summer monsoon rainfall anomalies in East Asia and the SM index, WY index, and EAP index, respectively. Comparing Fig. 6c with Figs. 6a and 6b, it may be clearly seen that the correlation between the summer monsoon rainfall anomalies in East Asia and the EAP index proposed in this study is larger than the other correlations, especially in the positive correlation region of the lower reaches of the Huaihe River, and South Korea and Japan, and in the negative correlation region of the southeastern coast of China and around the Philippines. Thus we can conclude that the EAP index defined in this study can better describe the interannual variations of the East Asian summer monsoon.

## 4. Relationship between the EAP index and the summer surface air temperature in East Asia

Generally, the interannual variability of summer surface air temperature has an out of phase relationship with the variability of summer rainfall in the East Asian monsoon region. Therefore, it is also possible that there is a close relationship between the summer surface air temperature anomalies and the EAP index. Figure 7 is the distribution of the correlation coefficients between the summer surface air temperature anomalies in China and the EAP index. It may be seen from the figure that there are large positive correlations in the Yangtze River valley, the Huaihe River valley, North China, and Northeast China. The maximum correlation coefficient reaches 0.6 in Northeast China and the Korean Peninsula, which greatly exceeds the confidence level of 95%. That is to say, in a summer with high EAP index, the summer surface air temperature may be high and a hot summer may occur in these regions. For example, in the summers of 1961, 1978, and 1994, drought and hot summers occurred in the middle and lower reaches of the Yangtze River and the Huaihe River valley. Our results suggest that the EAP index proposed in this study can describe the interannual variations of summer surface air temperature not only in the Yangtze River valley and the Huaihe River valley, but also in North China and Northeast China.

**Table 1.** The correlation coefficients among the summer rainfall (YH-R) and surface air temperature anomalies (YH-T) in the Yangtze River valley and the Huaihe River valley, and the summer rainfall (K-R) and surface air temperature anomalies (K-T) in South Korea, and the EAP index for 42 summers.

| Parameters | $I_{\mathrm{EAP}}$ | YH-R | YH-T | K-R | K-T |
|---|---|---|---|---|---|
| $I_{\mathrm{EAP}}$ | 1.00 | | | | |
| YH-R | –0.59 | 1.00 | | | |
| YH-T | 0.54 | –0.68 | 1.00 | | |
| K-R | –0.37 | 0.33 | –0.25 | 1.00 | |
| K-T | 0.77 | –0.56 | 0.66 | –0.50 | 1.00 |

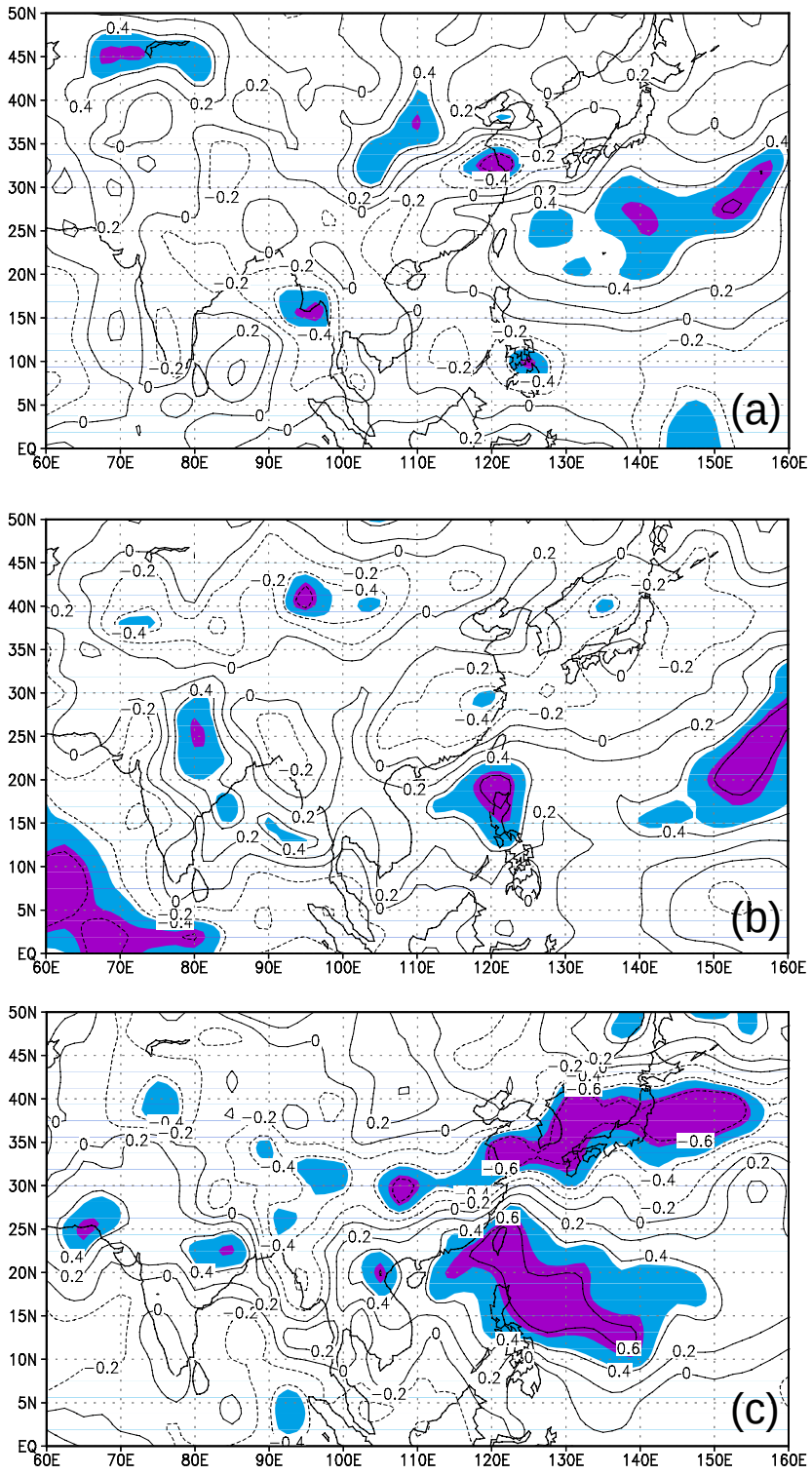



**Fig. 6.** Distribution of the correlation coefficients between the summer (June–August) rainfall anomalies in China with (a) the SM index, (b) the WY index, and (c) the EAP index. The shaded areas in the figure denote where the confidence level is over 95%.

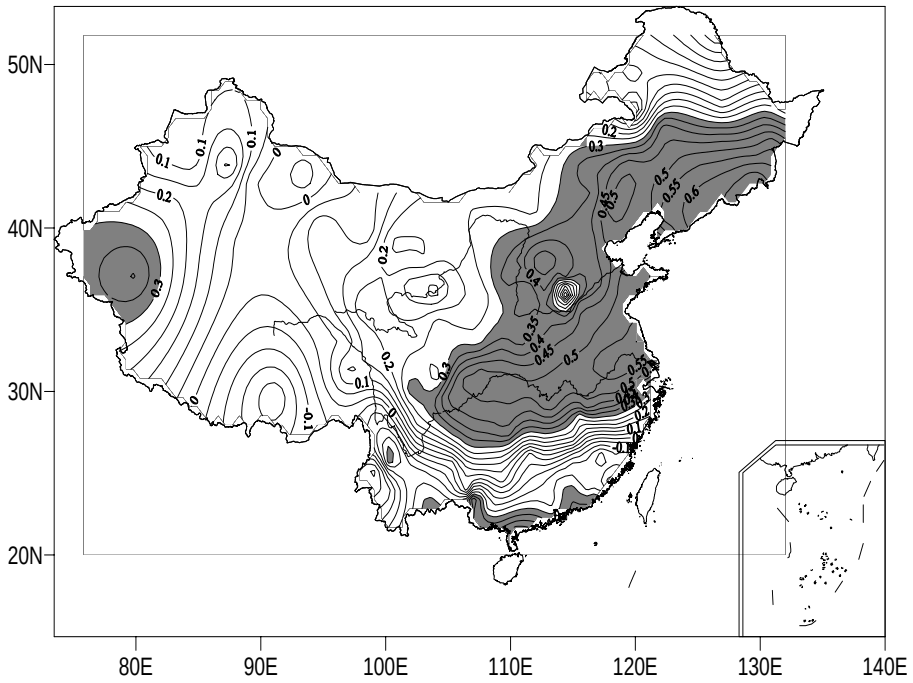



**Fig. 7.** Distribution of the correlation coefficients between the summer surface air temperature anomalies in China and the EAP index. The shaded areas denote where the confidence level is over 95%.

Similarly, the EAP index can also describe the interannual variations of summer surface air temperature anomalies in South Korea (figures not shown). Table 1 shows that the correlation between the summer surface air temperature anomalies in Korea and the EAP index is significant with a correlation coefficient of 0.77, which exceeds the confidence level of 99%.

From the above results and the correlation coefficients shown in Table 1, it is clearly seen that the EAP index based on the EAP teleconnection pattern can well describe not only the interannual variations of summer rainfall, but also the interannual variations of summer surface air temperature anomalies in the East Asian summer monsoon region, especially in the Yangtze River valley and the Huaihe River valley of China, and in South Korea and Japan. Therefore, the interannual variation of the EAP index defined in this study can be used to denote the interannual variability of monsoon climate in East Asia.

## 5. Relationship between the EAP index and the East Asian summer monsoon circulation

As mentioned above, there is a close relationship between the EAP index and the interannual variations of summer rainfall and surface air temperature in East Asia. But what causes the close relationship between the EAP index and the interannual variability of monsoon climate in East Asia? This may be due to the fact that the EAP index can well indicate the interannual variability of the East Asian summer monsoon circulation. In order to demonstrate this, the composite anomalous distributions of the 500-hPa height fields, the wind fields and their meridional component at 700 hPa over East Asia for the summers with high EAP index, (i.e., $I_{\rm EAP} \geqslant 1$), namely, 1961, 1978, and 1994 and for the summers with low EAP index, ($I_{\rm EAP} \leqslant -1$), namely, 1954, 1957, 1980, 1987, 1996, and 1998, are respectively analyzed.

Figures 8a and 8b indicate the composite distributions of the 500-hPa height anomaly fields over East Asia for the summers with $I_{\rm EAP} \geqslant 1$ and for the summers with $I_{\rm EAP} \leqslant -1$, respectively. Comparing Fig. 8a with Fig. 8b, it may be seen that the composite anomaly pattern of summer monsoon circulations for the summers with high EAP index is opposite to those for the summers with low EAP index. In a summer with high EAP index, as shown in Fig. 8a, the western Pacific subtropical high shifts northward. The positive 500-hPa height anomaly area is located over the Huaihe River valley, North China and the southern part of Northeast China, the Korean Peninsula and Japan, and the negative 500-hPa height anomaly areas are located over South China and eastern Siberia. Therefore, in a summer with high EAP index, the

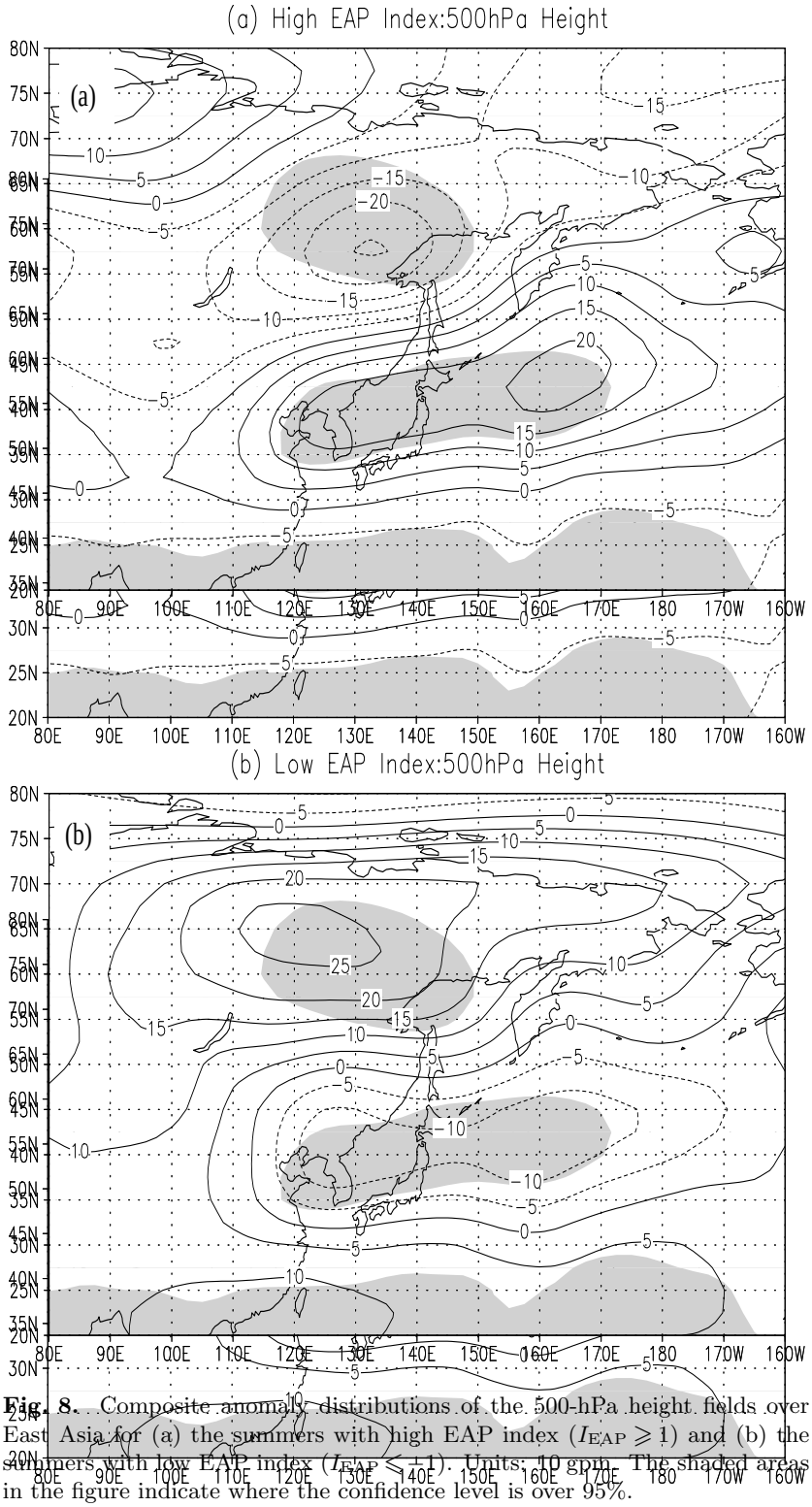



**Fig. 8.** Composite anomaly distributions of the 500-hPa height fields over East Asia for (a) the summers with high EAP index ($I_{\mathrm{EAP}} \geqslant 1$) and (b) the summers with low EAP index ($I_{\mathrm{EAP}} \leqslant -1$). Units: 10 gpm. The shaded areas in the figure indicate where the confidence level is over 95%.

summer surface air temperature is high and the summer monsoon rainfall is less in the middle and lower reaches of the Yangtze River and the Huaihe River valley, Korea, and Japan. In contrast, in a summer with low EAP index, as shown in Fig. 8b, the western Pacific subtropical high shifts southward. The negative 500-hPa height anomaly area is located over the Huaihe River valley, North China and Northeast China, the Korean Peninsula and Japan, and the positive 500-hPa height anomaly areas are located over South China and eastern Siberia. Therefore, in a summer with low EAP index, the summer surface air temperature is normal or below normal, and the summer monsoon rainfall is strong in the middle and lower reaches of the Yangtze River and the Huaihe River valley, Korea, and Japan.

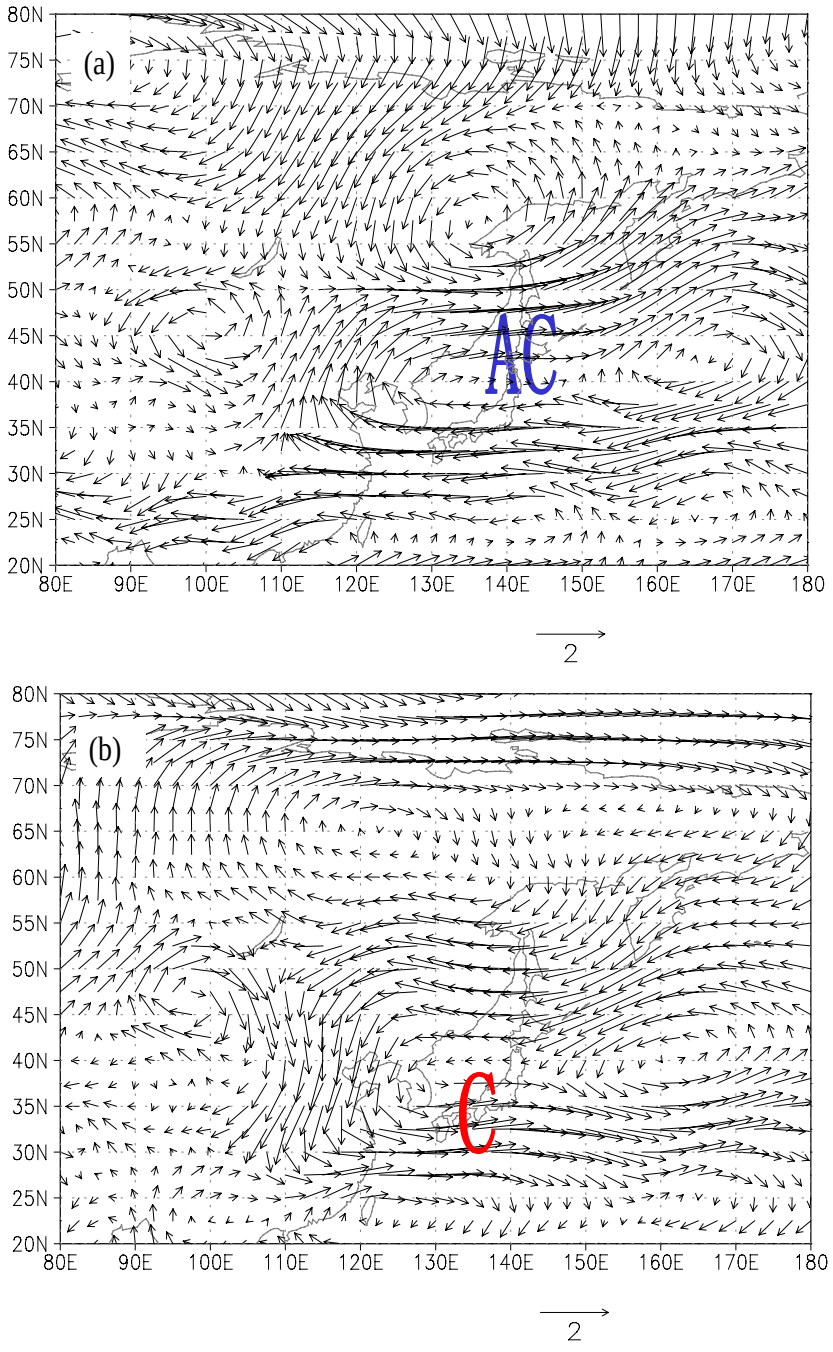



**Fig. 9.** Same as Fig. 8 but for the horizontal wind anomaly distribution at 700 hPa.

Generally, the monsoon flow is most obvious at 700 hPa over East Asia. Thus, the composite anomaly distributions of the wind fields at 700 hPa over East Asia for the summers with high EAP index ($I_{\rm EAP} \geqslant 1$), namely, 1961, 1978, and 1994, and for the summers with low EAP index ($I_{\rm EAP} \leqslant -1$), namely, 1954, 1957, 1980, 1987, 1996, and 1998, are respectively analyzed (see Fig. 9). As shown in Fig. 9a, in a summer with high EAP index, the strong anomalous anticyclonic circulation is located over the Huaihe River valley, North China, and Northeast China, and the anomalous cyclonic circulation is located over South China and eastern Siberia. In contrast, in a summer with low EAP index, as shown in Fig. 9b, the anomalous cyc-

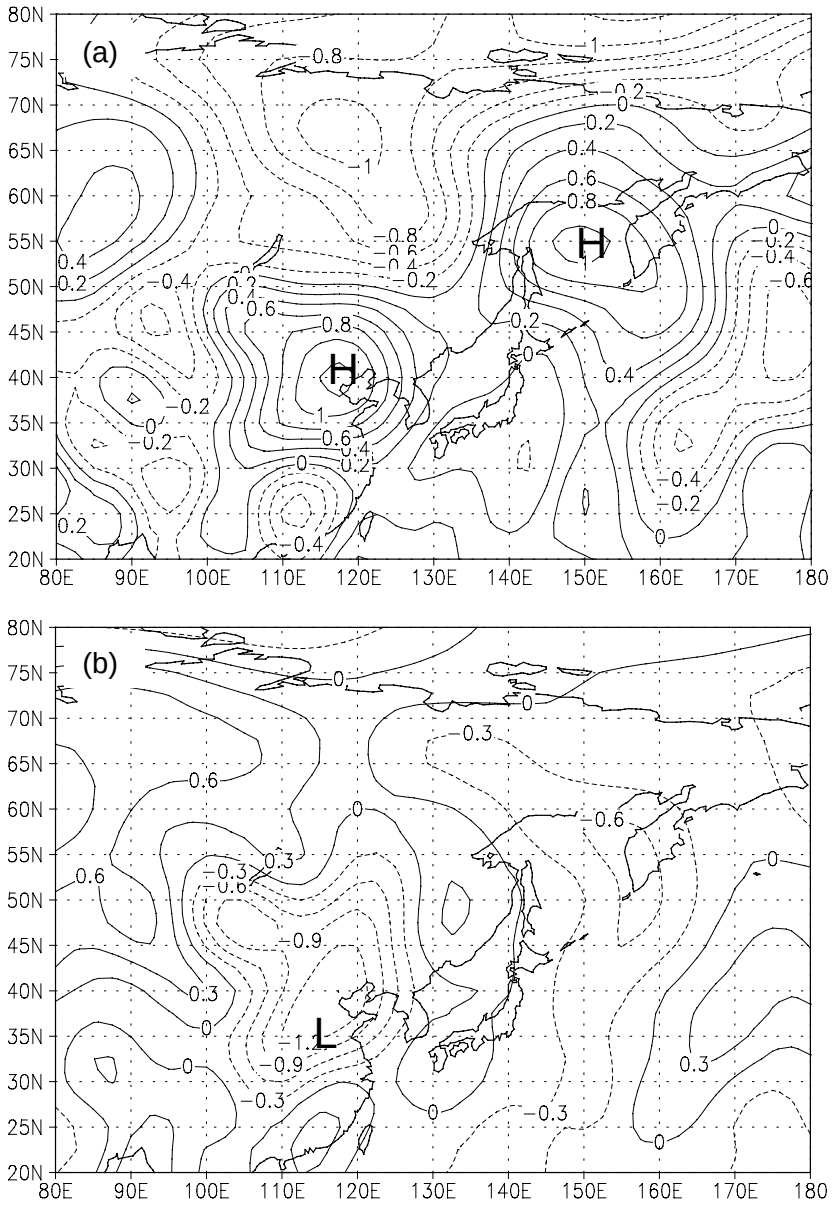



**Fig. 10.** Same as Fig. 8 but for the meridional wind anomalies at 700 hPa. Units: m $s^{-1}$.

lonic circulation is located over the Huaihe River valley, North China and Northeast China, the Korean Peninsula, and Japan, and the anomalous anticyclonic circulations are located over South China and eastern Siberia. Therefore, the anomaly circulation pattern shown in Fig. 9a is opposite to that shown in Fig. 9b. These patterns are similar to the distributions of the 500-hPa height anomaly field shown in Fig. 8.

From the above analyses, it is clear that the EAP index can well describe the interannual variability of summer monsoon circulation over East Asia.

An important characteristic of the summer monsoon flow over East Asia is its large meridional component. Huang et al. (1998) pointed out that the meridional component of water vapor transport associated with the East Asian summer monsoon is dominant. The convergence of water vapor transport, which can cause rainfall, is closely associated with the meridional transport of water vapor from south to north by monsoon flow. Therefore, the composite distributions of the meridional component of the wind fields at 700 hPa over East Asia for the summers with high EAP index and for the summers with low EAP index are also analyzed and shown in Figs. 10a and 10b, respec-

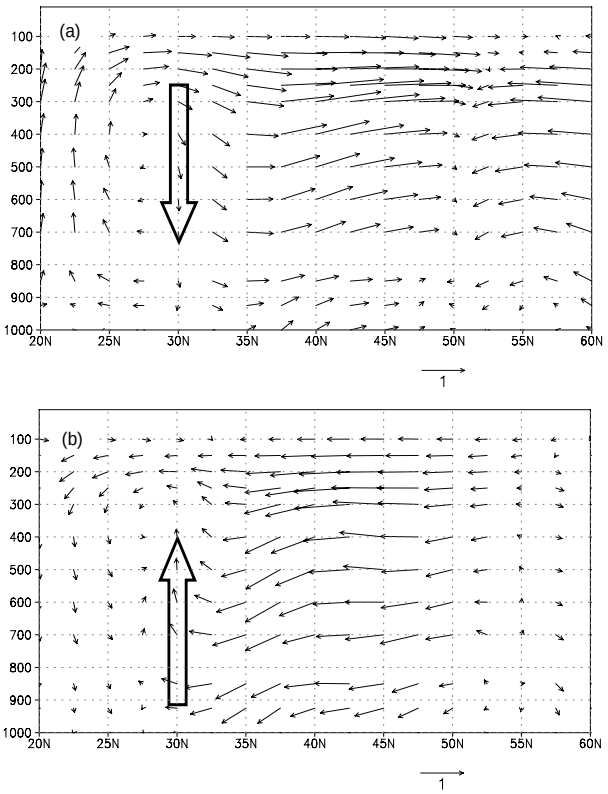



**Fig. 11.** Same as Fig. 8 but for the vertical-meridional circulation anomalies averaged between 110°–125°E.

tively. Comparing Fig. 10a with Fig. 10b, the anomalous meridional component of the wind fields at 700 hPa for the summers with high EAP index is different from that for the summers with low EAP index. In a summer with high EAP index, the positive meridional wind anomalies appear over the Huaihe River valley, North China, and the Korean Peninsula, i.e., the strong southerly winds of the monsoon flow over these regions. However, in a summer with low EAP index, the negative meridional wind anomalies appear over the Huaihe River valley, North China, and the Korean Peninsula, as shown in Fig. 10b. This indicates that in a summer with low EAP index, the southerly monsoon flow is weak over these regions.

## 6. Relationship between the EAP index and the vertical-meridional circulation over East Asia

Tao and Chen (1987) pointed out that the mean vertical-meridional circulation over the East Asian summer monsoon region is different from that over the Indian monsoon region. In East Asia there is no high mountain barriers and the flow can frequently penetrate from the mid-latitudes into the subtropical region of East Asia. In the climatological mean state, the convective activities are strong over the monsoon trough (or ITCZ) located at about 15°N. The outflow in the upper troposphere caused by the convective activities can be directed equatorward and northward from 15°N. Another ascending flow can be caused by the strong convective activities over the mei-yu front in the Yangtze River valley and the Baiu front in Japan. Thus the northward flow from the monsoon trough converges with the southward outflow due to the strong convective activities over the mei-yu front. This convergence can form a strong descending flow, which contributes to the formation of the western Pacific subtropical high.

Here the results show that the vertical-meridional circulation over the East Asian summer monsoon region in a summer with high EAP index is very different from that in a summer with low EAP index. Figures 11a and 11b are the composite anomaly distributions of vertical-meridional circulation averaged along 110°–125°E for the summers with $I_{\mathrm{EAP}} \geqslant 1$ and for the summers with $I_{\mathrm{EAP}} \leqslant -1$, respectively. Comparing Fig. 11a with Fig. 11b, it may be found that the vertical-meridional circulation in a summer with high EAP index is opposite to that in a summer with low EAP index. As shown in Fig. 11a, the positive anomaly distribution of vertical-meridional circulation, i.e., the Hadley cell-like circulation, is located over the region from South China to the Yangtze River valley. And there are strong southerly wind anomalies not only in the lower troposphere but also in the upper troposphere over the area to the north of 30°N. In this case, there is strong descending anomaly flow at 30°N. This means that the ascending flow is not strong over the mei-yu front in China and the Baiu front in Japan. Thus, the monsoon rainfall and the convective activities are not strong in these regions. However, there are strong southerly wind anomalies over the Yellow River valley, North China, and North Korea, which may cause strong rainfall. In a summer with low EAP index, as shown in Fig. 11b, the anomaly distribution of vertical-meridional circulation exhibits an anti-Hadley cell-like circulation, or a monsoon cell-like circulation, over the region from South China to the Yangtze River valley. There are strong northerly wind anomalies in both the lower and upper troposphere over the area to the north of 30°N. In this case, there is a strong ascending anomaly flow at 30°N, indicating a strong ascending flow over the mei-yu front in China and the Baiu front in Japan. The monsoon rainfall and the convective activities are strong in these regions. However, the southerly winds are weak over the Yellow River valley, North China, and North Korea. And the monsoon rainfall is weak in North China and North Korea in this case.

Therefore, our results also show that the EAP index proposed in this study can describe well the interannual variability of vertical-meridional circulation over East Asia.

## 7. Conclusion and discussion

Based on the EAP (East Asia/Pacific) teleconnection pattern in the summer atmospheric circulation anomalies over the Northern Hemisphere suggested by Nitta (1987), and Huang and Li (1987, 1988), respectively, an index measuring the strength of the East Asian summer monsoon is defined in this paper. This index is called the EAP index. The composite analyses of observed data clearly show that the EAP index can well describe not only the interannual variations of summer rainfall and surface air temperature in the Yangtze River valley and the Huaihe River valley, the Korean Peninsula, and Japan, but also the interannual variability of the East Asian summer monsoon system including the horizontal monsoon circulation and the vertical-meridional circulation cell over East Asia.

With the NCEP/NCAR reanalysis data and summer rainfall and surface air temperature data from the observational stations of China and Korea, it is shown that in a summer with high EAP index, the monsoon rainfall is weak in the Yangtze River valley and the Huaihe River valley, South Korea, and Japan. And the surface air temperature is also high in the Huaihe River valley, North China and Northeast China, and North Korea. In this case, the anticyclonic anomaly circulation is located over the Yangtze River valley and the Huaihe River valley, Korea, and Japan. The southerly monsoon flow is strong over East Asia. A positive vertical-meridional anomaly circulation cell is located over the area from South China to the Yangtze River valley, and there is a descending anomaly flow at about 30°N. However, in a summer with low EAP index, the monsoon rainfall is strong in the Yangtze River valley and the Huaihe River valley, South Korea, and Japan. The surface air temperature is low in the Huaihe River valley, North China and Northwest China, and North Korea. In this case, the cyclonic anomaly circulation is located over the Yangtze River valley and the Huaihe River valley, Korea, and Japan. The southerly monsoon flow is weak over East Asia. A reverse vertical-meridional anomaly circulation cell is located over the area from South China to the Yangtze River valley, and there is an ascending anomaly flow at about 30°N.

The analyses of observed data show that the EAP index can well denote the interannual variations of summer monsoon climate in East Asia because the index can well describe the East Asian summer monsoon circulation system. However, our study is preliminary. Further research is needed to improve the definition of this index so that it can be more suitable for measuring the interannual variations of the East Asian summer monsoon.

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# Indian Ocean Capacitor Effect on Indo–Western Pacific Climate during the Summer following El Niño

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ABSTRACT

Significant climate anomalies persist through the summer (June–August) after El Niño dissipates in spring over the equatorial Pacific. They include the tropical Indian Ocean (TIO) sea surface temperature (SST) warming, increased tropical tropospheric temperature, an anomalous anticyclone over the subtropical northwest Pacific, and increased mei-yu–baiu rainfall over East Asia. The cause of these lingering El Niño effects during summer is investigated using observations and an atmospheric general circulation model (GCM). The results herein indicate that the TIO warming acts like a capacitor anchoring atmospheric anomalies over the Indo–western Pacific Oceans. It causes tropospheric temperature to increase by a moist-adiabatic adjustment in deep convection, emanating a baroclinic Kelvin wave into the Pacific. In the northwest Pacific, this equatorial Kelvin wave induces northeasterly surface wind anomalies, and the resultant divergence in the subtropics triggers suppressed convection and the anomalous anticyclone. The GCM results support this Kelvin wave–induced Ekman divergence mechanism. In response to a prescribed SST increase over the TIO, the model simulates the Kelvin wave with low pressure on the equator as well as suppressed convection and the anomalous anticyclone over the subtropical northwest Pacific. An additional experiment further indicates that the north Indian Ocean warming is most important for the Kelvin wave and northwest Pacific anticyclone, a result corroborated by observations.

These results have important implications for the predictability of Indo–western Pacific summer climate: the spatial distribution and magnitude of the TIO warming, rather than simply whether there is an El Niño in the preceding winter, affect summer climate anomalies over the Indo–western Pacific and East Asia.

## 1. Introduction

El Niño–Southern Oscillation (ENSO) is the dominant mode of climate variability on instrumental records. Typically, El Niño develops during boreal summer, peaks during early winter, and decays the following spring (Fig. 1). Hereafter, seasons refer to those for the Northern Hemisphere. We denote the ENSO-developing year as year 0 and the following year as year 1. Thus, the summer [June–August (JJA)] of the ENSO-developing year is symbolized as JJA(0), while the summer of the following year as JJA(1). By JJA(1), eastern Pacific sea surface temperature (SST) returns either to or slightly below normal. By shifting tropical convection, El Niño excites atmospheric teleconnection into other tropical oceans and the extratropical Pacific–North America (PNA) sector. The so-called atmospheric bridge effect in the tropics is manifested in the delayed spring warming of the Indian (Klein et al. 1999) and North Atlantic (Enfield and Mayer 1997) Oceans. Because of large SST anomalies in the tropics and an effective tropics-to-midlatitude waveguide, the PNA teleconnection is most pronounced during the peak phase of El Niño, providing winter climate predictability for North America (Trenberth et al. 1998; Alexander et al. 2002).

Pronounced ENSO-related anomalies linger mysteriously during JJA(1) after eastern Pacific SST anomalies have dissipated. Chinese researchers have long suggested that during the summer, following El Niño,

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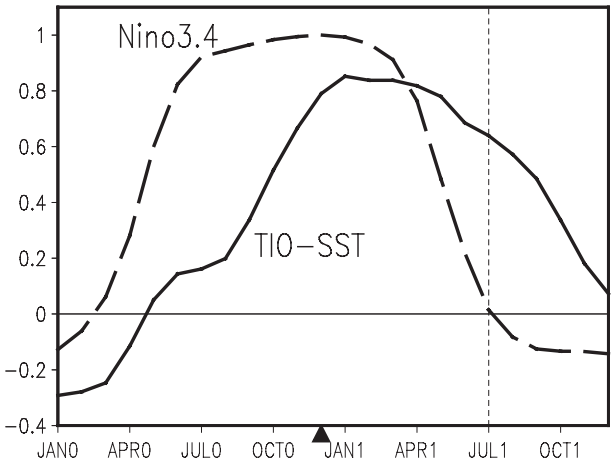



FIG. 1. Correlation of tropical Indian Ocean (20°S–20°N, 40°–100°E) SST (solid) with the Niño-3.4 (5°S–5°N, 170°–120°W) SST index for November(0)–December(0)–January(1). Numerals in parentheses denote years relative to El Niño: 0 for its developing and 1 for decay year. The dashed curve is the Niño-3.4 SST autocorrelation as a function of lag. The black triangle denotes December(0), the peak phase of ENSO.

pronounced precipitation anomalies tend to take place near the Yangtze River (Fu and Teng 1988; Huang and Wu 1989; Huang et al. 2004), causing major floods there as in 1954 and 1998 summers. Over the northwest (NW) Pacific, climatic anomalies can be even larger and more robust during the summer following than preceding the El Niño peak, despite much weaker SST anomalies in the eastern Pacific.

Figure 2 illustrates some of the mysterious JJA(1) anomalies based on surface observations over the NW Pacific, as represented by correlation with the Niño-3.4 (5°S–5°N, 120°–170°W) SST index at its peak phase [November–December (ND)(0)–January(J)(1), hereafter NDJ(0)]. At Guam (13.5°N, 145°E), precipitation is not significantly correlated with ENSO during its developing phase. A negative correlation begins to develop around December–February (DJF)(1), peaks during February–March(1), and persists through JJA(1) (Fig. 2a, solid) when the eastern Pacific SST index already turns slightly negative. Consistent with the rainfall correlation, sea level pressure (SLP) in the broad subtropical NW Pacific begins to develop strong positive correlation with El Niño at NDJ(0), which peaks during the spring (Wang et al. 2003). This correlation decays thereafter but remains significant at the 95% level through JJA(1) (Fig. 2b). The South China Sea (SCS) is the third example, displaying pronounced warming twice at the peak phase of El Niño (Liu et al. 2004) and July–September (JAS)(1) (Xie et al. 2003; Wang et al. 2006), respectively (Fig. 2c). The summer warming is associated with a reduction in the southwest monsoon and ocean upwelling it induces. The delayed ENSO effects extend to the midlatitude atmosphere. Precipitation at Hachijo Island (33.1°N, 139.8°E) south of Tokyo, Japan, displays two peaks in correlation with ENSO (Fig. 2a, dashed): one in DJF(1) and one in JJA(1), both of which are well above the 95% significance level. The summer correlation at Hachijo is of the opposite sign to that at Guam, linked by a meridional teleconnection pattern (Nitta 1987). Our last example is Ishigaki Island (24.3°N, 124.2°E), Japan, where sea level shows a significant positive correlation at a very long lag, during November–December(1) (Fig. 2d). Such prolonged ENSO effects during JJA(1) are the focus of the present study.

Over the subtropical NW Pacific, an anomalous high pressure system with an anticyclonic surface circulation begins to appear at the ENSO mature phase (winter) and develops into maximum intensity during spring (Harrison and Larkin 1996; Wang et al. 2000, 2003), broadly consistent with Fig. 2b. At its peak phase (spring), this surface high is roughly collocated with negative SST anomalies, suggestive of a local ocean–atmosphere interaction (Wang et al. 2000; Lau and Nath 2003). Wang et al. (2000) propose that SST-induced anticyclonic circulation, superimposed on the spring northeast trades, acts to reinforce the initial SST anomalies via surface evaporation and wind stirring. Their mechanism is a variant of wind–evaporation–SST (WES) feedback that emphasizes the role of meridional wind instead of zonal wind in the previous literature (Xie 2004). The anticyclonic anomaly persists into JJA(1) as corroborated by in situ observations (Fig. 2), but the local SST feedback does not seem to persist through summer. A local feedback requires positive correlation between SST and precipitation but observations indicate that this correlation is weak or even negative during summer (Wang et al. 2005). Results from the present study show that atmospheric anomalies during JJA(1) over the NW Pacific are forced not by local SST, but remotely by Indian Ocean SST.

The tropical Indian Ocean (TIO) warms up following the El Niño peak because of the downwelling ocean Rossby waves in the southwestern basin (Xie et al. 2002; Huang and Kinter 2002) and heat flux changes elsewhere (Klein et al. 1999; Tokinaga and Tanimoto 2004). See Schott et al. (2008) for a recent review of the TIO response to ENSO. During the El Niño mature phase, precipitation decreases over the TIO despite this basin-wide warming, leading to a notion that the TIO warming is largely a passive response to ENSO teleconnection without much climatic influences. Countering this notion, however, recent modeling studies suggest that the TIO warming is partially responsible for the formation of the

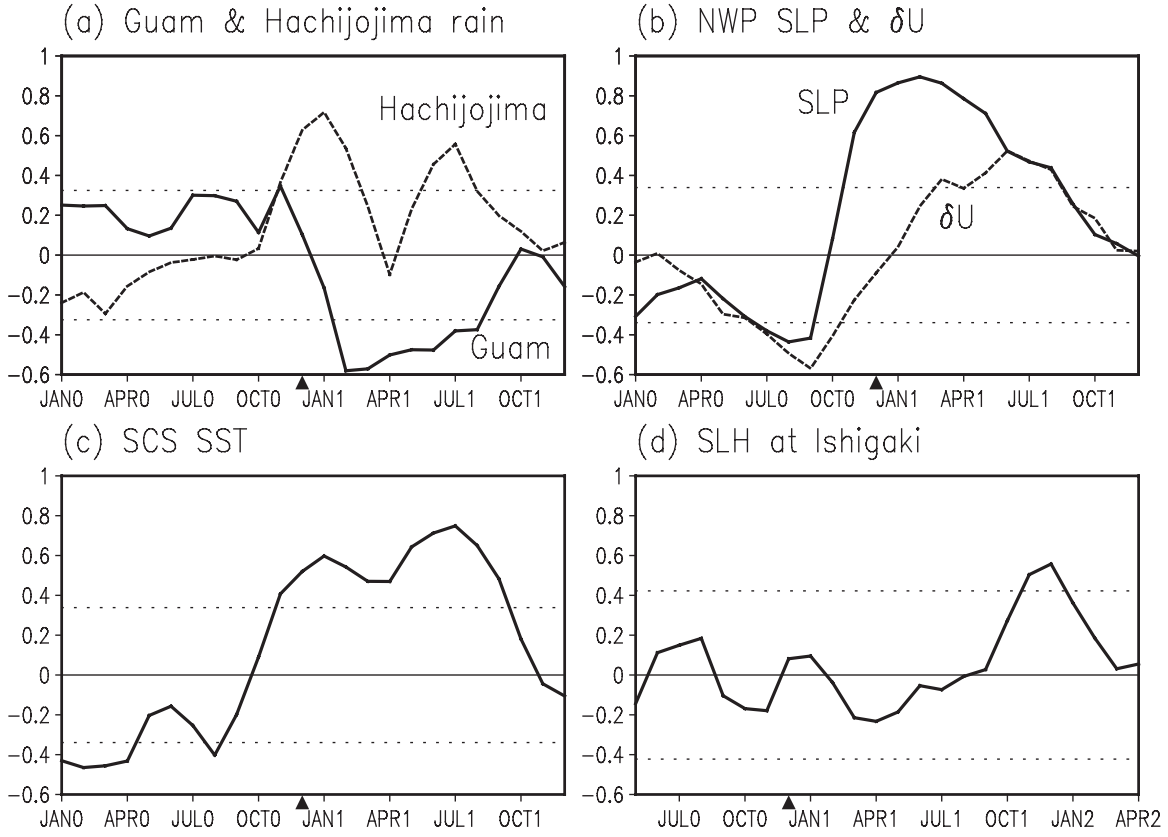



FIG. 2. Correlation with the NDJ(0) Niño-3.4 SST index: (a) rainfall at Guam and Hachijo Islands; (b) subtropical NW Pacific SLP (10°–25°N, 130°–170°E), and surface wind vorticity defined as (27°–33°N) minus (11°–15°N) zonal wind difference averaged in 120°–150°E; (c) SCS SST (5°–15°N, 110°–125°E); and (d) sea level height at Ishigaki Island. (b) and (c) Based on ICOADS. Dotted lines indicate the 95% significance level, which varies with the data record length. (In (a), the significance level is 0.44 for the shorter Hachijo record.) The black triangle denotes December(0), the peak phase of ENSO.

NW Pacific anticyclone anomaly from winter to spring (Watanabe and Jin 2003; Annamalai et al. 2005), and its persistence into JJA(1) (Yang et al. 2007). These results form the following capacitor hypothesis (Yang et al. 2007): El Niño teleconnection causes TIO SST to increase like a battery charging a capacitor; the TIO warming persists through the summer [JJA(1)], and exerts its climatic influences on surrounding regions after the El Niño decay, like a discharging capacitor. Early studies based on correlation (Wu et al. 1995) and idealized general circulation model (GCM) experiments (Wu and Liu 1995) have hinted at TIO SST effects on East Asian-western Pacific climate in summer. More recently, Yoo et al. (2006) note that spring TIO SST is a good predictor for the summer NW Pacific anticyclone. While they do not clarify what maintains this relationship between spring SST and summer atmospheric anomalies, Fig. 1 shows that the TIO SST persistence is the likely cause. Ohba and Ueda (2006) suggest that SST gradients between the North Indian and NW Pacific Oceans are important for June precipitation variability east of the Philippines. While the persistence of the NW Pacific anticyclone through summer may indeed be due to the TIO capacitor effect, the mechanism by which TIO SST affects the subtropical NW Pacific remains unclear.

The TIO capacitor effect may go beyond the subtropical NW Pacific surface circulation. Tropospheric temperature is known to increase over the tropics following El Niño (Yulaeva and Wallace 1994; Sobel et al. 2002), and the tropospheric warming lasts much longer than El Niño itself. Atmospheric GCM experiments show that much of the persistence in zonal-mean tropospheric temperature increase is not due to eastern Pacific SST, but to the SST warming over the tropical Indian and Atlantic Oceans (Kumar and Hoerling 2003; Lau et al. 2005). It is unclear whether and how the lingering tropospheric warming is related to surface anomalies, such as the subtropical NW Pacific anticyclone.

The present study examines robust, ENSO-induced climatic anomalies during JJA(1) when SST anomalies have mostly vanished over the equatorial Pacific. They include the TIO warming, NW Pacific anticyclone,

and tropospheric warming. As reviewed briefly above, these anomalies have previously been studied somewhat in isolation. We note that most literature on the NW Pacific anticyclone deal with its formation and development from winter to spring (Wang et al. 2000, 2003; Watanabe and Jin 2003; Annamalai et al. 2005), but few focus on its maintenance for JJA(1) (Yang et al. 2007). Summer is the rainy season for East Asia when precipitation variability has huge socioeconomic impacts. The present study attempts a synthesis to reveal relationships among and mechanisms for these anomalies by tracking ENSO correlation into JJA(1). Specific questions of interest include the following: What maintains the robust precipitation and circulation anomalies over the NW Pacific? Is there evidence for remote influences of the TIO on the NW Pacific; what about local SST effect? Is the long-lasting tropospheric warming in any way related to surface anticyclone over the NW Pacific? Observational diagnostics are the main approach but numerical modeling is also used as necessary to test hypotheses.

In the following, section 2 describes the datasets and atmospheric GCM. Section 3 presents observational diagnosis, discusses the ENSO-induced anomalies during JJA(1) in the global tropics and over the Indo–western Pacific region, and tests the local SST and TIO capacitor hypotheses. Section 4 examines atmospheric anomalies in response to the TIO warming in models. Section 5 is a summary and discusses broad implications.

## 2. Methods

### *a. Observational data*

We use the Hadley Centre Global Sea Surface Temperature (HadSST) dataset (Rayner et al. 2006), the Center for Climate Prediction (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1996), and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) atmospheric reanalysis (Kalnay et al. 1996), originally on 1°, 2.5°, and 2.5° grids, respectively. We have regridded these datasets onto a common 2.5° latitude × 10° longitude grid for a 29-yr period from 1979 to 2007 (limited by CMAP's satellite rainfall estimates). All of the features to be discussed have large zonal scales, and the coarse zonal grid spacing is to help suppress small-scale variability.

We complement the above data-assimilated/statistically interpolated datasets with ship observations in the International Comprehensive Ocean–Atmospheric Data Set (ICOADS; Worley et al. 2005), originally on a 2° grid. Grid points without at least one observation for a given month are flagged as missing. No additional interpolation is made. We regrid ICOADS onto the same 2.5° latitude × 10° longitude grid, and limit our analysis to after 1950, when observations are relatively abundant.

To extend the precipitation record back beyond the CMAP period, monthly observations at Guam are used to track rainfall over the subtropical and midlatitude NW Pacific, respectively. We average data at five stations to form an island-mean precipitation series, which runs nearly continuously from 1945. A different method by normalizing data at each station with its standard deviation yields a similar island-mean series. Observations at Hachijo Island for a shorter period (1980–2007) are used to monitor rainfall in the midlatitude NW Pacific. Monthly-mean sea level height data (1969–2007) at Ishigaki Island are obtained from the University of Hawaii Sea Level Center.

The present study focuses on interannual variability associated with ENSO. To reduce the effect of pronounced intraseasonal variability over the Indo–western Pacific Oceans, we perform a 3-month running average. A 9-yr running mean is then applied (separately for each calendar month) to remove decadal and longer variations, which are significant over the tropical Indo-Pacific Oceans (Deser et al. 2004; Du and Xie 2008). We use SST averaged over the eastern equatorial Pacific (Niño-3.4: 5°S–5°N, 120°–170°W) to track ENSO, which is referred to as the ENSO index. Typical decorrelation time for such bandpass-filtered time series is slightly more than half a year (Fig. 1). We estimate the total degree of freedom as (analysis period in years)/1.5. For a 29-yr time series, a correlation of 0.43 reaches the 95% significance level based on $t$ test.

### *b. GCM*

The atmospheric GCM used is the ECHAM5, the latest Hamburg version of the European Centre for Medium-Range Weather Forecasts (ECMWF) model. A detail description of ECHAM5 is given by Roeckner et al. (2003). ECHAM5 employs a spectral dynamic core. We use a version with triangular truncation at zonal wavenumber 63 (T63; equivalent to 1.9° horizontal resolution) and 19 sigma levels in the vertical.

The model is forced with the observed monthly climatology of SST and sea ice. We analyze a 20-yr period of the control (CTL) simulation. We carry out two additional experiments. In the TIO run, we add 1°C SST over the TIO (20°S–20°N, 40°–110°E). In the other experiment, the same SST anomalies are imposed on the South Indian Ocean (SIO; 20°S–0°, 40°–110°E). Experiments with a more realistic 0.5°C SST anomaly yield similar results. In both TIO and SIO runs, the SST

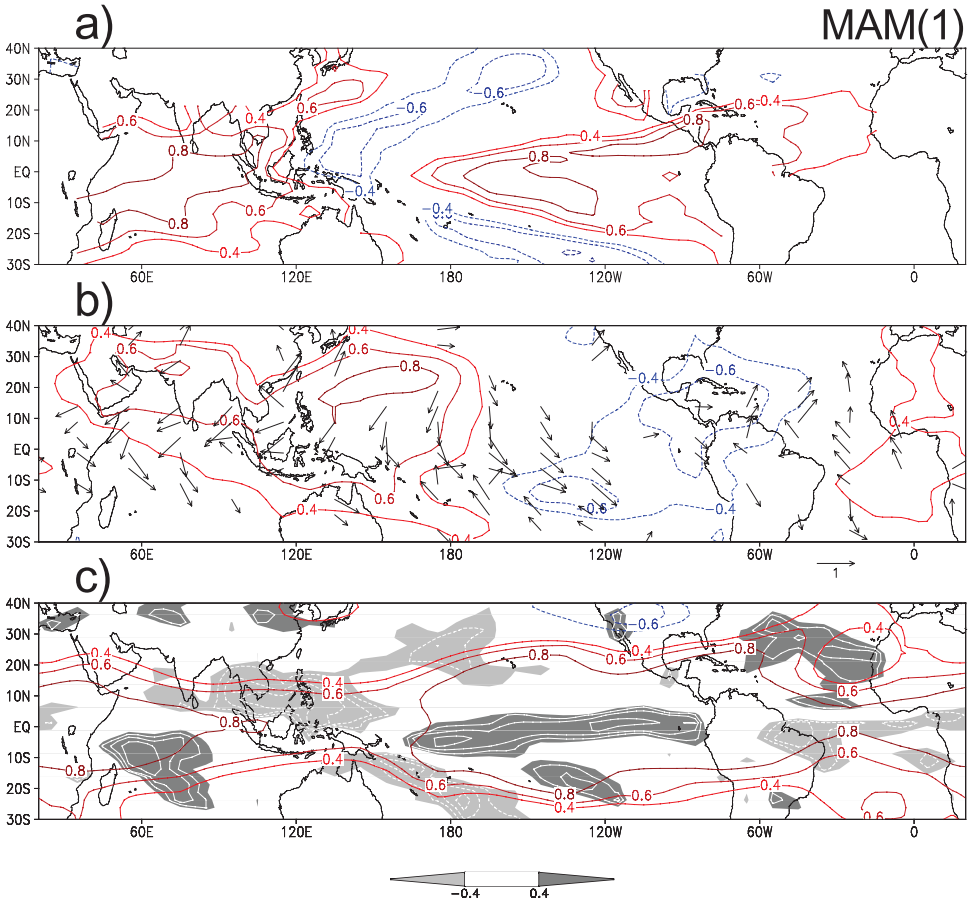



FIG. 3. MAM(1) correlation with the NDJ(0) Niño-3.4 SST index: (a) SST, (b) SLP (contours) and surface wind velocity, and (c) precipitation (gray shade and white contours at intervals of 0.1) and tropospheric temperature (contours).

anomalies are kept constant in time and the model is integrated for 20 yr. Thus, the experiments are equivalent to 20-member ensemble runs.

## 3. Observational analysis

This section examines spatiotemporal variations in ENSO correlation to infer the mechanisms for atmospheric anomalies.

### *a. MAM(1)*

During March–May (MAM)(1), El Niño begins to decay but large positive anomalies of SST and precipitation remain in the eastern equatorial Pacific (Fig. 3). The westerly wind anomalies near the international date line are displaced south of the equator, rendering them ineffective to sustain the Bjerknes feedback and El Niño (Vecchi and Harrison 2006; Lengaigne et al. 2006). Precipitation decreases over the off-equatorial western Pacific, with a broad high pressure anomaly over the NW Pacific. The anomalous winds there intensify the prevailing northeast trades, contributing to the SST cooling (Wang et al. 2000). The upwelling Rossby waves in the ocean may also be a factor for the surface cooling (Wang et al. 1999).

Positive SST anomalies cover the entire TIO, with the maximum (>0.8) over the tropical southwest Indian Ocean (SWIO), a signature of the downwelling Rossby waves over the shallow thermocline dome (Murtugudde and Busalacchi 1999; Behera et al. 2000; Xie et al. 2002; Huang and Kinter 2002). While the SST anomalies are more or less symmetrical, precipitation anomalies are nearly antisymmetric about the equator, positive over the SWIO, and negative over the eastern north Indian Ocean (NIO). Surface winds show a C-shaped pattern typical of the response to antisymmetric heating, northeasterly north and northwesterly south of the equator. Kawamura et al. (2001) and Wu et al. (2008) suggest that this antisymmetric atmospheric pattern is due to be the WES feedback

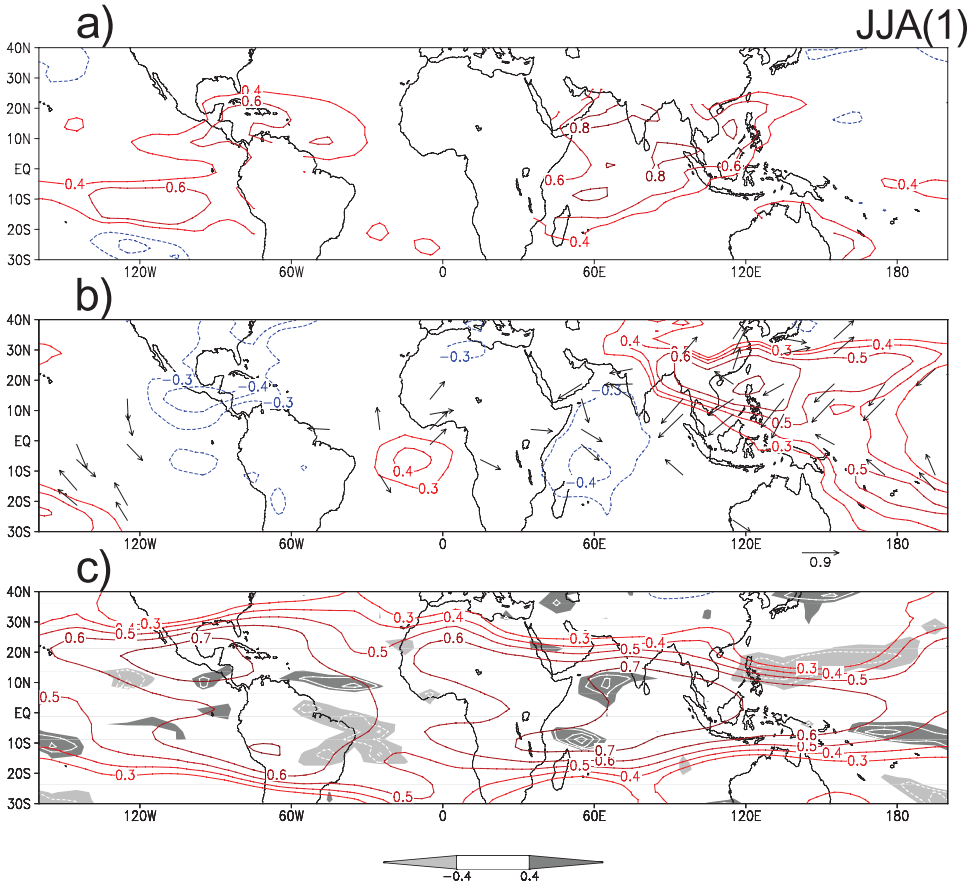



FIG. 4. Same for Fig. 3, except for JJA(1).

within the TIO. The northeasterly anomalies, in particular, help sustain the NIO warming through summer after the onset of the southwest monsoon in May (Du et al. 2009).

The free troposphere warms up along the entire equatorial belt, with tropospheric temperature correlation above 0.8 everywhere except over the western Pacific (minimum ~0.5° at 160°E). The correlation features an off-equatorial maximum on either side of the equator over the central/eastern Pacific and an equatorial maximum to the east, reminiscent of the Matsuno–Gill pattern (Matsuno 1966; Gill 1980) anchored by the eastern Pacific warming. This tropical-wide tropospheric warming takes place despite the tropical-mean precipitation anomaly being nearly zero, supporting the idea that deep convection adjusts tropospheric temperature to a moist-adiabatic profile over the warming eastern Pacific, spreading the tropospheric warming throughout the tropics via fast equatorial waves (Su et al. 2001; Chiang and Sobel 2002). El Niño–induced tropospheric warming has been proposed to cause SST to increase over other tropical oceans that support deep convection (Chiang and Sobel 2002; Chiang and Lintner 2005).

### b. JJA(1)

Figure 4 shows JJA(1) correlations with the NDJ(0) ENSO index. SST anomalies vanish along the equator across the Pacific while SST correlation remains strong and exceeds 0.6 over much of the TIO. In contrast to MAM(1), when the decaying El Niño remains the major forcing for atmospheric anomalies, strong atmospheric correlations with coherent large-scale structure shift to the TIO and neighboring NW Pacific. The strongest atmospheric anomalies at the surface are located over the NW Pacific, including decreased rainfall and increased SLP. An anomalous anticyclone develops there, with northeasterly anomalies from the Bay of Bengal to the international date line. This summer anticyclone has been noted in previous studies (e.g., Wang et al. 2003) and the resultant weakening of the southwesterly monsoon wind causes a pronounced warming over the SCS (Xie et al. 2003; Wang et al. 2006). SST correlation is weakly negative but statistically marginal over the

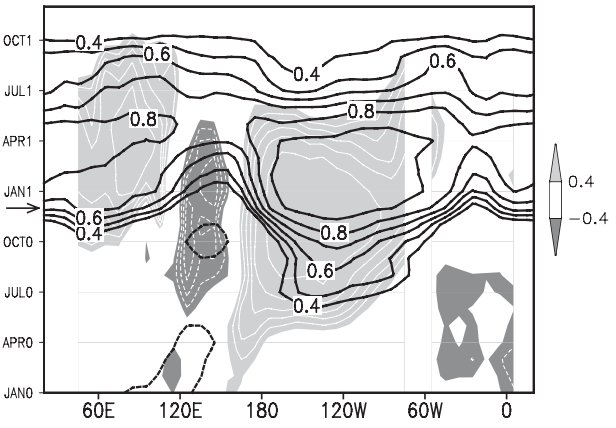



FIG. 5. Longitude–time section of correlation with the NDJ(0) Niño-3.4 index: SST (light shade $> 0.4$; dark $< -0.4$; white contours at intervals of 0.1), and tropospheric temperature (black contours) on the equator. The black arrow denotes December(0), the peak phase of ENSO.

eastern half of the NW Pacific anticyclone (maximum correlation $r_{\max} \sim -0.4$), suggesting that local SST is not the main anchor for the decreased precipitation and increased SLP there. (More analysis of the SST correlation will be presented in section 3c.) This anticyclone is the most significant surface atmospheric anomaly during JJA(1) ($r_{\max} \sim 0.7$) that originates from ENSO, while the TIO warming is the most significant major SST anomaly ($r_{\max} \sim 0.8$), implying that the latter is the cause of the former. An empirical orthogonal function (EOF) analysis confirms this conclusion (see the appendix).

Tropospheric temperature displays a Matsuno–Gill pattern that is consistent with a localized heating in the TIO, with a Kelvin wave–like wedge penetrating into the western Pacific on the equator and Rossby wave–like, off-equatorial maxima over the western Indian Ocean to Africa. Tropical convection adjusts tropospheric temperature close to a moist-adiabatic profile, consistent with the equivalent potential temperature ($\theta$e) in the planetary boundary layer (Emanuel et al. 1997). Thus, an SST increase over a convective region acts to warm the tropospheric column. The spatial pattern of tropospheric temperature anomalies (Figs. 3c, 4c) appears to support this moist-adiabatic adjustment to ocean warming. Figure 5 shows tropospheric temperature correlation with the NDJ(0) ENSO index in the tropical Pacific and TIO, along with the SST correlation. Over the tropical Pacific, the troposphere begins to warm up soon after the development of El Niño. The tropospheric warming there reaches the maximum about a season after the El Niño peak and decays rapidly after April(1), falling below the 95% significance level by July(1). Over the TIO, on the other hand, tropospheric warming commences quite late, developing rapidly in November(0) and reaching the peak in April(1). It persists at high correlation until August(1), and then decays rapidly with the TIO SST warming.

SST correlation falls to near zero on the equator (as the thermocline begins to shoal), but significant SST warming remains on either side of the equator in the eastern Pacific during JJA(1) (Fig. 4a). The southern SST warming does not have much of a signature in SLP because the mean SST is too cold for deep convection (Figs. 4b,c). North of the equator, by contrast, the SST warming over the far eastern Pacific warm pool and Caribbean Sea induces weak negative SLP anomalies and appears to reinvigorate the off-equatorial tropospheric warming. Because of the great geographical distance, it is unlikely that the lingering eastern Pacific SST anomalies north of the equator are the main cause of the NW Pacific anticyclone.

### c. *Kelvin wave–induced divergence off the equator*

Precipitation correlation is weakly positive over the warm TIO, but not well organized in space. Restoring tropospheric temperature toward a moist-adiabatic profile, Su and Neelin (2003) show that their model simulates tropospheric temperature and precipitation anomalies during the developing and mature phases of El Niño. They report that the precipitation response is rather complicated and not necessarily correlated with local SST.

Given weak local SST correlation, the robust NW Pacific anticyclone at the surface is mysterious. We propose a mechanism for the TIO to force this surface anticyclone remotely. The moist-adiabatic adjustment provides a mechanism that couples SST and tropospheric temperature over the warm TIO (Emanuel et al. 1997; Neelin and Su 2005). Via this coupling, the TIO warming forces a warm equatorial Kelvin wave to the east. Figure 6 superimposes tropospheric temperature and surface wind correlations with the NDJ(0) ENSO index. Surface friction drives northeasterly winds onto the equatorial low pressure in the baroclinic Kelvin wave, inducing surface divergence and suppressing deep convection in the subtropical NW Pacific.

The circulation–convection feedback is important to amplify this Kelvin wave–induced Ekman divergence (WIED). While the Kelvin wave in the free troposphere is nearly symmetric about the equator, the response of convective and surface anomalies clearly favors the summer hemisphere (Fig. 6), where convective feedback is strong with high SST and precipitation in the mean. The negative rainfall anomalies

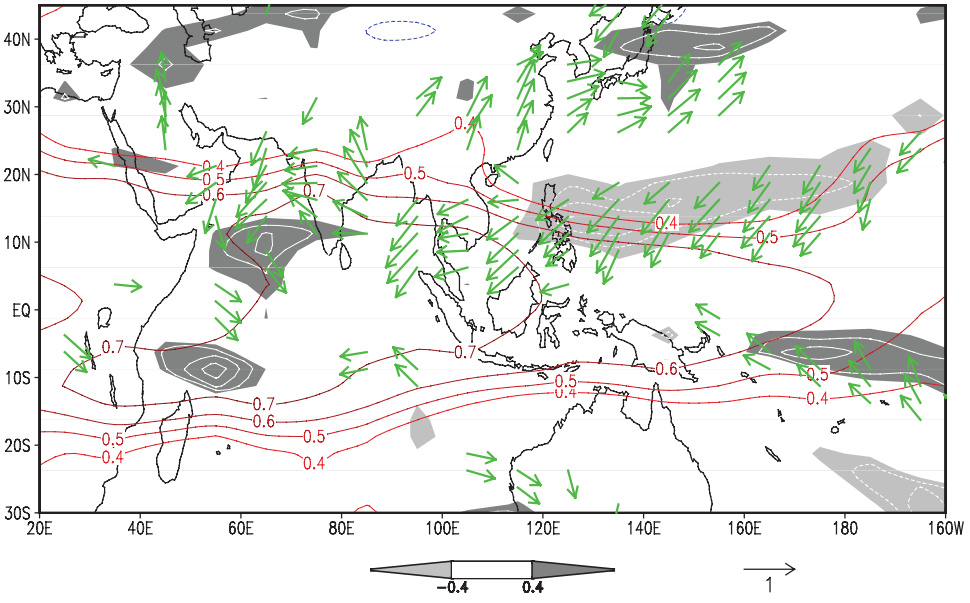



FIG. 6. JJA(1) correlation with the NDJ(0) Niño-3.4 SST index: tropospheric (850–250 hPa) temperature (contours), precipitation (white contours at intervals of 0.1; dark shade > 0.4; light < −0.4), and surface wind velocity (vectors).

over the NW Pacific excite an equatorial Kelvin wave of an opposite sign, weakening the warm Kelvin wave wedge emanating from the TIO. Indeed, the warm Kelvin wave wedge is much shorter in zonal extension than the off-equatorial Rossby waves to the west (Fig. 4c), opposite to the Matsuno–Gill model prediction.[1] In the warm Kelvin wave wedge near the equator, precipitation anomalies are not very coherent as the deep subsidence of the wave in the free troposphere opposes the surface convergence in creating precipitation changes. Section 4 tests this hypothesis further with an atmospheric GCM.

Modeling studies suggest a TIO contribution to the NW Pacific anticyclone anomaly during MAM(1) (Su et al. 2001; Watanabe and Jin 2003; Annamalai et al. 2005). We note that from the NIO to NW Pacific during MAM(1), northeasterly wind anomalies and the negative precipitation band are located on the northern edge of the warm Kelvin wave wedge (Fig. 3c). This collocation of surface anomalies with the Kelvin wave's northern flank is consistent with our Kelvin WIED mechanism.

### *d. Northwest Pacific precipitation*

The NW Pacific warm pool reaches far north during summer, with the 27°C SST contour expanding north of 30°N (Fig. 7). In addition to the basin-scale intertropical convergence zone (ITCZ) anchored by an off-equatorial warm SST band, a subtropical rainband develops during late summer and early fall (July–September) in the NW Pacific, riding on the expanding warm pool. This subtropical rainband appears as a local meridional maximum in precipitation (15°–25°N), and it is a key element of the NW Pacific monsoon (Murakami and Matsumoto 1994; Wang and LinHo 2002). The late-July onset of this subtropical NW Pacific convection weakens and displaces the mei-yu–baiu rainband northward that covers east China and Japan during June to mid-July (Ueda et al. 1995, 2009).

Interestingly, summer precipitation variance is much stronger in the subtropical rainband than in the ITCZ over the NW Pacific despite the mean precipitation being much larger in the latter (Fig. 7). Figure 8a shows the time–latitude section of local correlation between monthly SST and precipitation over the western Pacific. In the subtropical NW Pacific, the subtropical rainfall variance intensifies during July–September but is not correlated with the local SST. Our analysis of ENSO correlation indicates that much of the subtropical NW Pacific rainfall variability is instead remotely forced, by the TIO warming in particular. This TIO capacitor effect is pronounced during JJA(1), possibly corresponding to interannual variability in the onset and early development period of subtropical NW Pacific convection. The cause of high variance from August onward needs further investigation.

[1] Factors such as the mean flow and its shear as well as heating latitude also influence the apparent zonal extent of these waves.

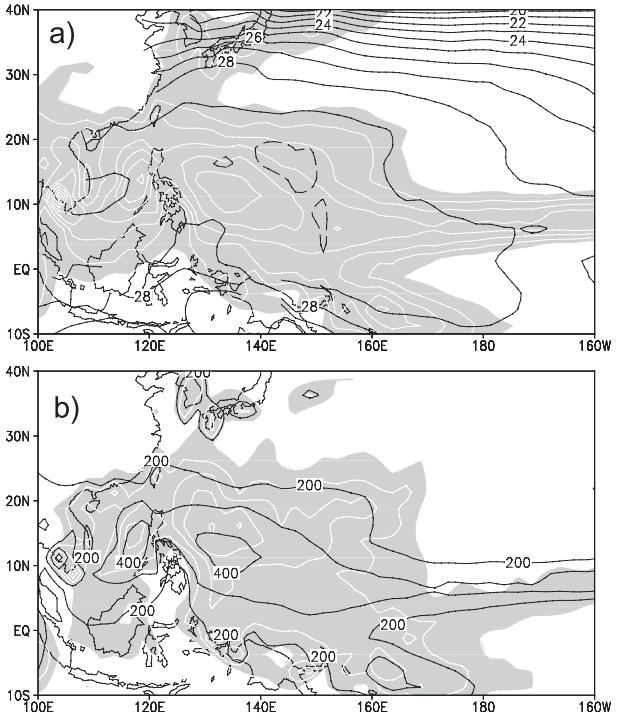



FIG. 7. (a) July–September climatology of SST (black contours with 29.5°C dashed) and precipitation (shade > 150 mm month$^{-1}$; white contours at 50 mm month$^{-1}$ intervals). (b) Standard deviation of precipitation for July–September (shade > 75 mm month$^{-1}$; white contours at 25 mm month$^{-1}$ intervals), along with the precipitation climatology (black contours).

The anticyclone anomaly over the subtropical NW Pacific begins to develop during the El Niño winter. Figure 9 shows the ENSO correlations of precipitation (shade), SST (color contours), and surface wind velocity in the NW Pacific. In Fig. 9, the onset of the El Niño–induced anticyclone in December(0) is manifested as the sudden appearance of the anomalous northerlies and SST cooling in 0°–10°N. Wang et al. (2000) suggest that the rapid onset of the anticyclone and SST cooling is due to their mutual interaction as follows. The SST cooling north of the equator reduces precipitation, exciting a westward-propagating Rossby wave. The resulting northerly wind anomalies intensify the mean northeast trades and amplify the initial SST cooling via evaporation and wind stirring. Hereafter we call this the WES-V feedback to emphasize the importance of meridional wind perturbations, where the traditional WES feedback considers the zonal wind effect on evaporation (Xie and Philander 1994). Mean northerly winds ($V$) are a necessary condition for this WES-V feedback, which is indeed met during December–April (Fig. 8b). (One may use $V = 2$ m s$^{-1}$ as a threshold for the WES-V feedback to operate given that weather-related variance is high in the ITCZ.) The mean northerlies weaken from February on, and the WES-V mechanism is unlikely to be important from April into the summer, as corroborated by the rapid transition of the local SST–precipitation correlation from positive to negative in 0°–10°N during April–July (Fig. 8a). Here we have used the SST–precipitation correlation as a test for the ocean-to-atmospheric feedback.

Figure 9 shows that ENSO correlations also display a rapid transition from May(1) to July(1). The center of negative rainfall correlation is confined to 0°–10°N up to May(1), and jumps to 10°–20°N in June(1). The northward shift is also seen in the center of the anticyclonic vorticity (as measured by the meridional shear). The northward shift in the anticyclonic center suggests that different mechanisms are at work before May(1) and during JJA(1), consistent with our inference that Pacific and TIO SST anomalies are important for the respective periods. The northward shift of precipitation correlation is also consistent with the evolution of precipitation variance, whose maximum shifts from the ITCZ during winter and spring to the subtropical rainband during summer. The El Niño–induced negative SST anomalies in the ITCZ decay rapidly from April(1). At 145°E, the negative JJA(1) precipitation anomalies at 15°N are not associated with significant SST correlation (Fig. 9), indicating the importance of remote forcing for the maintenance of the subtropical anticyclone during JJA(1).

### e. Further teleconnections

Atmospheric anomalies in the subtropical NW Pacific during JJA(1) excite additional teleconnections in both the ocean and atmosphere. In the ocean, the anticyclonic wind curl anomalies force downwelling Rossby waves that propagate westward (results not shown from an ocean reanalysis), leading to a rise in sea level at Ishigaki Island that peaks in the following December (Fig. 2d), almost a year after the mature phase of El Niño. In the atmosphere, the southwesterly anomalies on the western flank of this anomalous anticyclone transport moisture from the SCS, increasing summer rainfall over the mid-/lower reaches of the Yangtze River of China. Suppressed convection over the subtropical NW Pacific, through upper-tropospheric convergence, excites the Pacific–Japan (PJ) teleconnection (Nitta 1987), with a low pressure centered east of Japan (Fig. 6). Developing under the vertically sheared climatological meridional flow, the PJ vorticity anomalies exhibit an apparent poleward phase tilt with height (Kosaka and Nakamura 2006). The surface low around Japan is associated with an anomalous increase in local precipitation near Japan as observed at Hachijo Island (Fig. 2a).

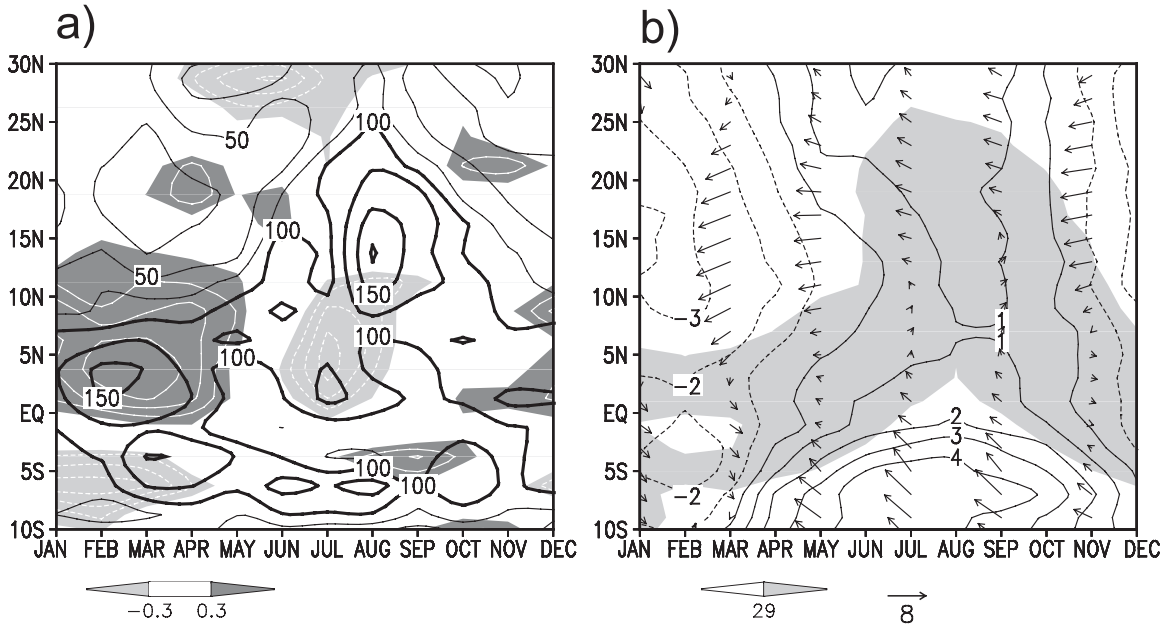



FIG. 8. (a) Monthly precipitation standard deviation (black contours, mm month$^{-1}$) and SST–precipitation correlation (dark shade $> 0.3$; light $< -0.3$; white contours at 0.1 intervals) at 145°E as a function of time and latitude. (b) ICOADS climatology of surface wind velocity (vectors, m s$^{-1}$), its meridional component (contours), and regions with SST $> 29$°C (shade) as a function of time and latitude over the western Pacific (145°E).

## 4. Model results

This section uses atmospheric GCM experiments to test the hypothesis that the TIO warming affects NW Pacific climate during summer. In the GCM, the imposed TIO warming[2] causes precipitation to increase over much of the basin (Fig. 10a). Over the equatorial Indian Ocean, precipitation somehow decreases despite the SST increase. Precipitation decreases and a surface anticyclone forms over the subtropical NW Pacific, with a cyclonic circulation to the northeast of Japan. In 0°–20°N from the NIO to western Pacific, the southwest monsoonal winds weaken, helping sustain the SST warming there in observations (Xie et al. 2003; Du et al. 2009). The model response is qualitatively consistent with observations (Figs. 4 and A1), but the NW Pacific anomalies are too weak while TIO precipitation anomalies too large. This result is consistent with a multimodel study of Li et al. (2008).

TIO-induced tropospheric warming forms a spatial pattern reminiscent of the Matsuno–Gill pattern, with the Kelvin wave response extending eastward and covering the entire equatorial Pacific (Fig. 10b). Compared with JJA(1) observations (Fig. 4), the eastward extension of the baroclinic Kelvin wave response is too large, resulting in unrealistic, strong surface easterly anomalies over the equatorial Pacific. This too-strong Kelvin wave response is related to too weak a precipitation reduction over the NW Pacific, which would otherwise excite a baroclinic Kelvin wave of the opposite sign and leaves little net anomalies in the equatorial Pacific. Thus, TIO warming indeed forces a baroclinic Kelvin wave with low SLP centered on the equator, triggering a precipitation reduction and surface anticyclone to the north. The model result of too weak a NW Pacific response suggests that the feedback between local convection and circulation is important for their anomalies to amplify and reach the observed magnitude. A comparison with the observed JJA climatology indicates that model precipitation is weaker by a factor of 2 over the subtropical NW Pacific, limiting the convection–circulation feedback. The poor simulations of subtropical NW Pacific summer rainfall have been flagged as a major problem for atmospheric GCMs (Kang et al. 2002).

The Japan pole of the PJ pattern is too weak and shifted northward compared to observations. Such deficiencies may be due to too-weak convective anomalies

[2] Care needs to be taken in designing SST-forced, atmospheric GCM experiments because SST changes may be a response to a remote forcing as happens over the TIO during the developing and mature phases of ENSO (Kumar and Hoerling 1998; Wu et al. 2006; Deser and Phillips 2006; Copsey et al. 2006). Our prescription of TIO SST as forcing here is justified by the facts that it is the most robust SST anomalies of the global ocean during JJA(1) and that tropospheric temperature anomalies are consistent with a Matsuno–Gill pattern forced by a heat source over the TIO (Fig. 4).

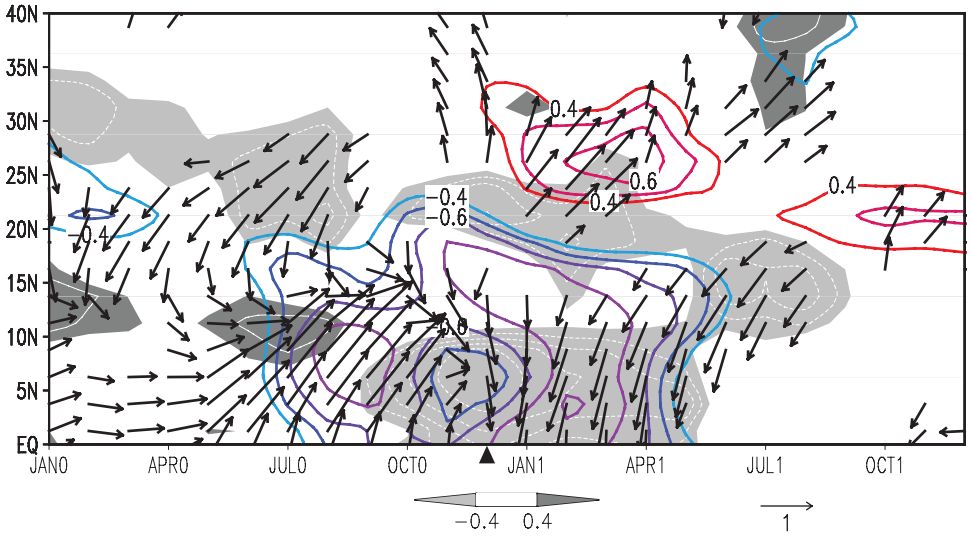



FIG. 9. Time–latitude section of correlation at 145°E with the NDJ(0) Niño-3.4 index: SST (color contours), precipitation (dark shade > 0.4; light shade < −0.4; white contours at 0.1 intervals), and surface wind velocity (vectors). The black triangle denotes December(0), the peak phase of ENSO.

in the subtropical NW Pacific and/or errors in simulating mean precipitation and mean flow, from which the PJ mode draws energy (Kosaka and Nakamura 2006).

### *a. Linear baroclinic model*

To illustrate the importance of the convection–circulation feedback, we turn to a primitive-equation model linearized around the observed JJA mean state

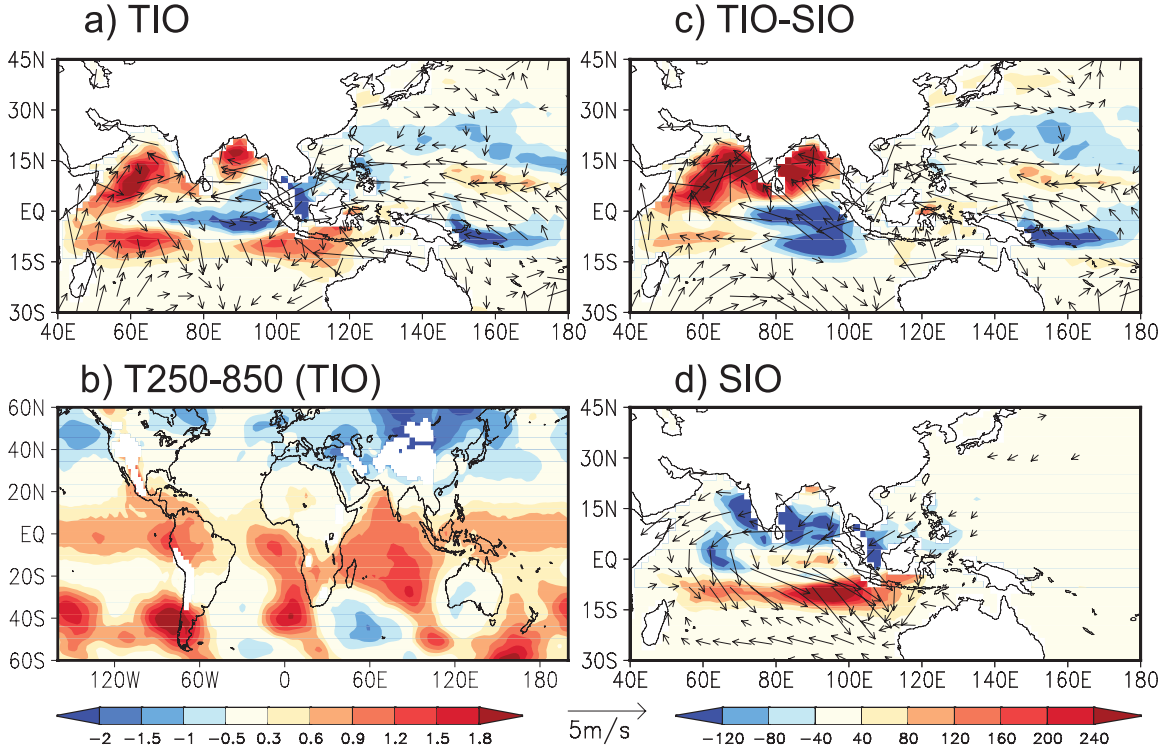



FIG. 10. Anomalies of precipitation (color, mm month$^{-1}$; right color bar) and surface wind velocity (m s$^{-1}$; wind speed smaller than 0.5 m s$^{-1}$ has been masked out) in (a) the TIO and (d) SIO atmospheric GCM runs, and (c) TIO–SIO difference. (b) Anomalies of tropospheric temperature (°C; left color bar) in the 250–850-hPa layer in the TIO run.

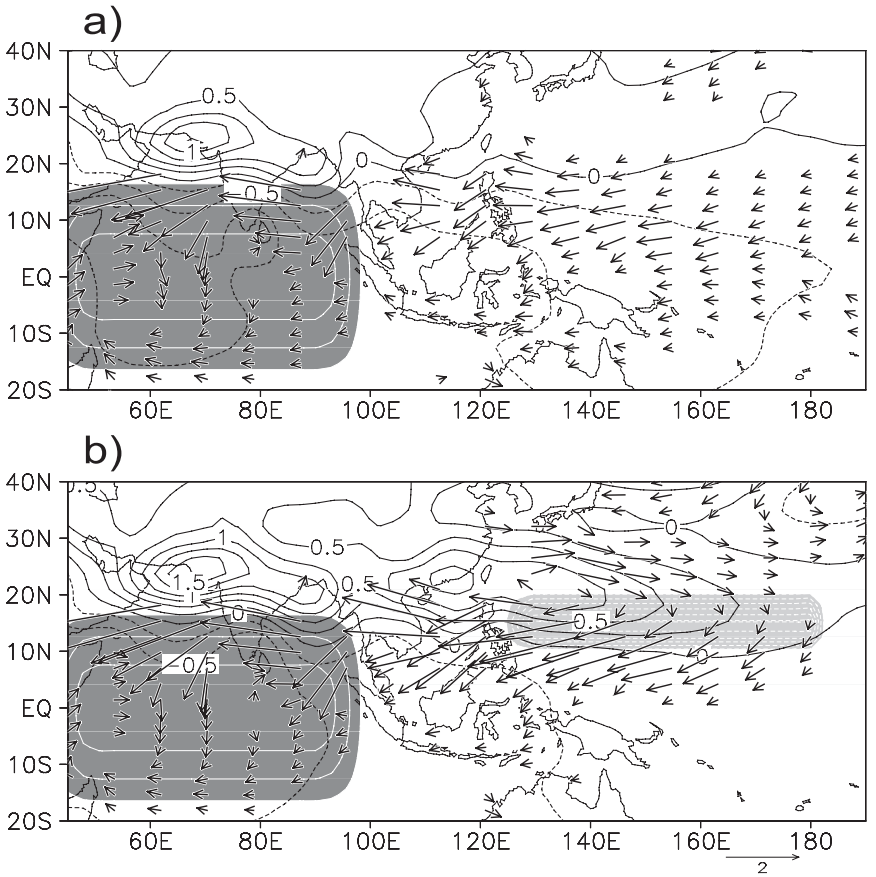



FIG. 11. The linear atmospheric model response to the TIO heating (a) with and (b) without interactive heating over the subtropical NW Pacific: sea level pressure (black contours, hPa), wind velocity (m $s^{-1}$) at 1000 hPa, and diabatic heating rate at 500 hPa (dark shade $> 0.25$; light $< -0.25$ K $day^{-1}$; white contours at 0.25 K $day^{-1}$ intervals).

as represented by the NCEP–NCAR reanalysis. The detailed description of the linear baroclinic model (LBM) may be found in Watanabe and Kimoto (2000). We use a version with T42 resolution in the horizontal and 20 sigma levels in the vertical. The model is forced by prescribed diabatic heating. In the vertical, the heating is confined between 900 and 100 hPa, with a peak at 500 hPa to mimic condensational heat release in deep convection. The heating ($Q$) is confined in the horizontal, following a half-cosine function in latitude ($y$),

$$Q = Q_0 \cos \pi\left[(y - y_s)/(y_n - y_s) - 1/2\right], \text{ for } y_s < y < y_n. \quad (1)$$

Within a longitudinal band $x_w < x < x_e$, it is zonally uniform except in a 5° zone on east and west sides where it ramps off to zero.

Figure 11a shows the LBM surface response near the steady state to a broad heating over the TIO: $x_w = 40°\text{E}$, $x_e = 100°\text{E}$; $y_s = 20°\text{S}$, $y_n = 20°\text{N}$. The maximum heating rate is set at 1 K $day^{-1}$. Because of the strong monsoon circulation in the JJA mean state, the low-level circulation anomalies are rather asymmetric, with a stronger response north than south of the equator. Equatorial symmetry strengthens at upper levels and near the international date line near the surface. The Kelvin wave response is visible in the sea level pressure field. On the northern flank of the Kelvin wave, surface friction turns the winds northeasterly, causing surface divergence there.

To test whether this Kelvin wave–induced divergence would lead to a strong anticyclone with the help of convection–circulation feedback, we conduct the following experiment with interactive heating over the NW Pacific. For simplicity, the horizontal and vertical structure of the interactive heating follows Eq. (1), with $x_w = 120°\text{E}$, $x_e = 180°$; $y_s = 10°\text{N}$, $y_n = 20°\text{N}$, a NW Pacific region chosen based on observations (Fig. 4). The magnitude of convective heating is parameterized as a linear function of surface convergence $Q_0 = -\alpha\nabla \bullet u_s$,

where $\mathbf{u}_s$ is the perturbation surface wind velocity averaged in the same NW Pacific region, and the coefficient $\alpha$ controls the strength of the convection–circulation feedback. Although crude, the empirically based parameterization is intended to illustrate the effect of this feedback on the development of NW Pacific anomalies in response to the TIO heating. The LBM is forced by the same broad TIO heating as before, but with an additional, interactive heat source over the NW Pacific. Convective feedback coefficient $\alpha$ is chosen so that the maximum heating rate is 1.8 K day$^{-1}$ in the steady state. The results are qualitatively insensitive to the choice of $\alpha$.

Figure 11b shows the LBM response on day 30 when it is in a nearly steady state. The model response is now dominated by a strong anticyclone from the SCS to NW Pacific, where the convection–circulation feedback amplifies the Kelvin wave–induced Ekman divergence and maintains a strong diabatic cooling. The model response resembles observed anomalies during JJA(1) from the north Indian Ocean to the NW Pacific (Figs. 4 and A1), including anomalous easterlies in 10°–20°N from the Arabian Sea to the international date line. The importance of reduced convection over the NW Pacific for the anticyclonic circulation is consistent with Terao and Kubota's (2005) linear model study using prescribed heating. The LBM with interactive NW Pacific heating even simulates a weak cyclonic circulation east of Japan, consistent with observations and with Kosaka and Nakamura's (2006) suggestion that the PJ pattern is a dynamical mode of the atmosphere shaped by the summer mean flow. The general agreement between the LBM experiment and observations supports our hypothesis that the Kelvin wave forced by the TIO warming triggers the suppression of convection and the development of a surface anticyclone over the NW Pacific.

### b. North Indian Ocean effect

An additional GCM experiment has been carried out to narrow down the TIO subdomain that is most important for the formation of the NW Pacific anticyclone. In the SIO run where the SST increase is limited to the tropical South Indian Ocean (SIO), precipitation increases south and decreases north of the equator, with a C-shaped, antisymmetric wind pattern (Fig. 10d), similar to observations over the western TIO (Fig. 6). The positive and negative anomalies of this precipitation dipole in the meridional direction are nearly equal in magnitude, rendering a small basin mean. As a result, anomalies over the Pacific are negligible. Figure 10c shows the TIO–SIO difference, which may be taken as the response to an NIO warming. The rainfall increase over the NIO is much larger than the reduction south of the equator. The net increase in the basin-mean precipitation forces a Kelvin wave response with low pressure and easterly wind anomalies over the equatorial Pacific, triggering a reduction in precipitation and the formation of a surface anticyclone over the subtropical NW Pacific. These model results indicate that SST anomalies over the NIO are more important for NW Pacific summer anomalies than those over the SIO. The preference for the NIO is likely due to higher mean SST, and hence stronger SST forcing of the atmosphere than over the SIO.

Figure 12 shows the observed SST correlation with a summer NW Pacific anticyclone index defined as the (27.5°–32.5°N) minus (10°–15°N) difference in surface zonal wind averaged in 130°–160°E. The correlation is low with local SST but strong over the NIO and SCS, corroborating the GCM results. SST correlation is also high over the equatorial southeast Indian Ocean, where negative IOD events sometimes develop during JJA(1), as in 1998. GCM results show that during summer of 1998, the east equatorial Indian Ocean warming is important in reducing precipitation anomalies in East Asian and the NW Pacific (Shen et al. 2001). There is a tendency for NW Pacific precipitation to correlate negatively with that over the TIO, the middle reach of the Yangtze River, and near Japan, broadly consistent with the GCM.

## 5. Summary and discussion

We have investigated ENSO-induced anomalies during the summer [JJA(1)] following its peak phase using observations and atmospheric models. The anticyclone centered in the subtropical NW Pacific emerges as the dominant anomaly pattern in the lower troposphere from our correlation and EOF analyses. This anomalous anticyclonic circulation explains many regional JJA(1) climate anomalies related to ENSO in Fig. 2. The weakened southwest monsoon causes the SCS to warm up; precipitation decreases at Guam and SLP increases in the anticyclone; the negative wind curls force downwelling ocean Rossby waves detected as a sea level rise at Ishigaki Island a few months later; and the PJ teleconnection from the subtropics intensifies baiu rainfall over Japan as observed at Hachijo Island.

Such summer atmospheric anomalies must be anchored by SST anomalies somewhere even though eastern Pacific SST anomalies have dissipated by JJA(1). The SST increase over the TIO and SCS is the most robust ocean anomaly pattern during JJA(1), making it a likely cause of persistent atmospheric anomalies (block arrow in Fig. 13). The tropospheric temperature anomaly pattern resembles the Matsuno–Gill response to a localized heat source over the TIO, offering additional support for the capacitor mechanism in which the El Niño–induced TIO warming

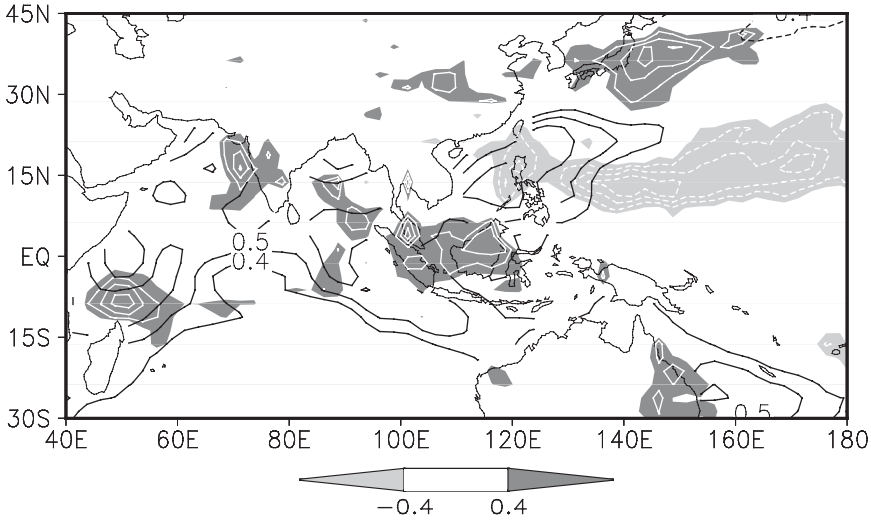



FIG. 12. Precipitation (gray shade and white contours) and SST (black contours) correlation during JJA with a NW Pacific anticyclone index, defined as (27.5°–32.5°N) minus (10°–15°N) zonal wind difference averaged in 120°–150°E. Only correlation above ±0.4 is shown, and contour intervals are 0.1.

persists through the summer and sustains atmospheric anomalies after El Niño itself is gone. Observed precipitation anomalies are weak over the TIO, suggesting the moist-adiabatic adjustment as the mechanism for the SST forcing of tropospheric temperature.

Atmospheric GCM experiments show that the long persistence of positive SST anomalies outside the tropical Pacific maintains the delayed tropospheric warming during and after the decay of El Niño (Kumar and Hoerling 2003; Lau et al. 2005). Our results suggest that this tropospheric warming triggers the suppression of NW Pacific convection. To our knowledge, our study is the first that relates these two robust atmospheric anomaly patterns that outlast El Niño. The JJA(1) tropospheric warming displays a Kelvin wave wedge extending into the western Pacific, acting to lower SLP near the equator and induce northeasterly winds and divergence north of the equator. The Ekman divergence initiates the suppression of deep convection over the NW Pacific (Fig. 6). The linear atmospheric model results support the Kelvin WIED mechanism and illustrate the importance of circulation–convection feedback for amplifying the surface high over the NW Pacific: surface divergence in the high pressure suppresses convection while the reduced latent heating intensifies the high and anticyclonic circulation.

We have conducted an atmospheric GCM experiment by increasing SST over the TIO. The model reproduces the baroclinic Kelvin wave over the equatorial Pacific and the anticyclone anomaly in the NW Pacific, in support of the above Kelvin wave–induced Ekman divergence mechanism. The NW Pacific anticyclone is too weak compared to observations because the model underestimates the summer precipitation climatology in the subtropical rainband, and hence the local circulation–convection feedback. An additional experiment with the SST warming confined to the south of the equator indicates that the NIO SST warming is the most important forcing of the Pacific response both on and north of the equator. This model result is consistent

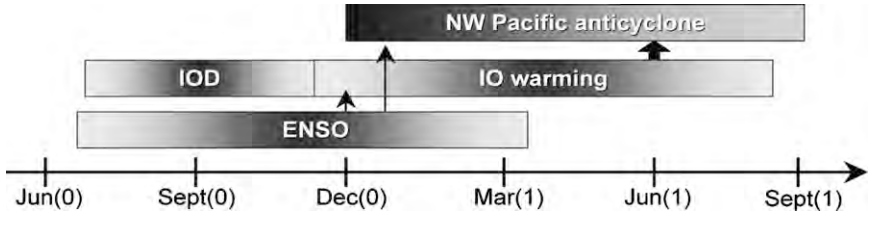



FIG. 13. Seasonality of major modes of Indo–western Pacific climate variability. Vertical arrows indicate causality, and the block arrow emphasizes the TIO capacitor effect, the major finding of the present study.

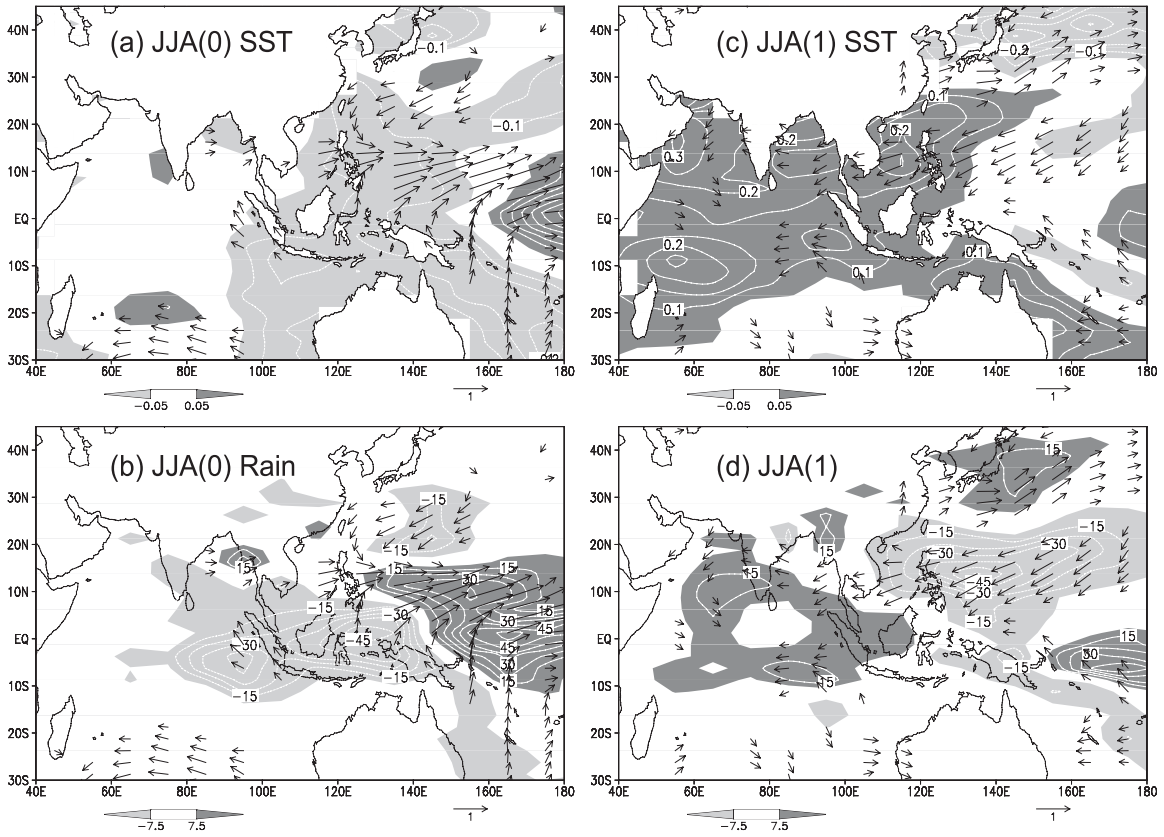



FIG. A1. Regression upon the NDJ(0) Niño-3.4 index for (left) JJA(0) and (right) JJA(1): (top) SST (°C) and (bottom) precipitation (mm month$^{-1}$).

with the observational analysis that NW Pacific anticyclone is most highly correlated with NIO SST.

The NW Pacific anticyclone anomaly begins to develop at the peak of El Niño and persists through the following summer (Harrison and Larkin 1996; Wang et al. 2003). The center of this anticyclone, as well as the center of negative precipitation anomalies, displays a northward shift from 5°–10°N during winter to 15°N during summer (Fig. 9). This meridional shift reflects a change in mechanism for maintaining the anticyclone. During winter, the precipitation decrease is collocated with negative SST anomalies, with the latter possibly being induced by the intensified northeasterly trades through the WES-V mechanism (Wang et al. 2000) with contributions from the TIO warming via the Kelvin WIED mechanism. During JJA(1), the precipitation decrease is most highly correlated with NIO SST, reflecting the remote influence from there, likely via the baroclinic Kelvin wave. Local JJA(1) SST anomalies are rather complicated, varying from positive in the western to marginally negative in the central and eastern parts of the NW Pacific anticyclone (Fig. 4).

The schematic in Fig. 13 summarizes the results. El Niño develops during the summer of year 0, peaks in winter, and decays in the spring of year 1. It induces the basin-wide warming over the TIO that persists through JJA(1). During the summer following El Niño, the TIO warming is the main culprit for the anomalous anticyclone and suppressed rainfall over the subtropical NW Pacific.

The following important question still remains: how is the TIO warming maintained through JJA(1)? A companion paper (Du et al. 2009) suggests that ocean–atmosphere interaction within the TIO is important.

TABLE A1. Statistics of summer (JJA) EOF modes over the Indo-western Pacific: explained variance (row 2), and season relative to the peak of El Niño (row 3) inferred from correlation (r) with the NDJ(0) ENSO index (row 4). For comparison, row 5 lists the simultaneous correlation with the JJA ENSO index. The PC pairs of (SST-2, Rain-1) and (SST-1, Rain-2) are correlated at 0.78 and 0.74, respectively.

| EOF | SST-2 | Rain-1 | SST-1 | Rain-2 |
|---|---|---|---|---|
| Variance (%) | 16 | 21 | 29 | 17 |
| Season inferred | JJA(0) | JJA(0) | JJA(1) | JJA(1) |
| r-Niño NDJ(0) | 0.71 | 0.76 | 0.69 | 0.75 |
| r-Niño JJA | 0.66 | 0.72 | 0.33 | 0.05 |

Specifically, the downwelling ocean Rossby wave anchors the warming, not only over the SWIO thermocline dome but also over the NIO by inducing northeasterly wind anomalies, which relax the southwest monsoon during early summer and reduce latent heat flux from the ocean.

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## APPENDIX

### JJA(0) and JJA(1) Comparison

A typical El Niño event affects the Indo–western Pacific climate during summers of both its developing and decay years. Figure A1 compares ENSO-induced anomalies between JJA(0) and JJA(1) based on a regression analysis referenced to the NDJ(0) ENSO index. During JJA(0), El Niño warming already takes shape over the equatorial Pacific, increasing rainfall and inducing westerly wind anomalies there to sustain the Bjerknes feedback. South of the equator from the Maritime Continent to the eastern Indian Ocean, SST decreases, suppressing atmospheric convection there. Over the eastern Indian Ocean, the SST cooling is coupled with the southeasterly wind anomalies, and the couplet often develops into an Indian Ocean dipole (IOD) mode, growing on the Bjerknes feedback (Saji et al. 1999; Webster et al. 1999; Schott et al. 2008). In the subtropical NW Pacific (15°–20°N), a surface cyclonic anomaly develops as part of the baroclinic Rossby wave forced by the Pacific warming.

While the centers of action for SST, precipitation, and surface winds are in the tropical Pacific during JJA(0), they shift to the Indo–western Pacific during JJA(1), anchored by the TIO warming (Figs. 4 and A1c,d). Decreased precipitation and the anticyclonic circulation over the subtropical NW Pacific are the most pronounced anomalies of the atmosphere during JJA(1). From the NIO to the NW Pacific, anomalies of precipitation and surface winds tend to have opposite signs during JJA(0) and JJA(1). This biennial tendency of the Indo–western Pacific response to ENSO, previously noted (e.g., Wang et al. 2003; Alexander et al. 2004), is due to the difference in forcing mechanism: the JJA(0) anomalies are largely forced directly by eastern Pacific SST while the JJA(1) anomalies are largely forced by TIO SST.

We have performed EOF analyses separately for precipitation and SST over the Indo–western Pacific Oceans (20°S–40°N, 40°E–180°) for the JJA season. The dominant modes are similar to the regression patterns in Fig. A1 (Table A1). Specifically, the second SST and first precipitation modes resemble the JJA(0) regressions, explaining 16% and 21% of the total variance, respectively. Their principal components (PCs) correlate highly with the NDJ(0) ENSO index. The first SST and second precipitation EOF modes resemble the JJA(1) regressions, explaining 29% and 17% of the total variance, respectively. Their PCs correlate highly with the ENSO index 6 months before [NDJ(0)], but not simultaneously [JJA(1)], which is indicative of the TIO capacitor effect.

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