

Diverging hydrological sensitivity among tropical basins

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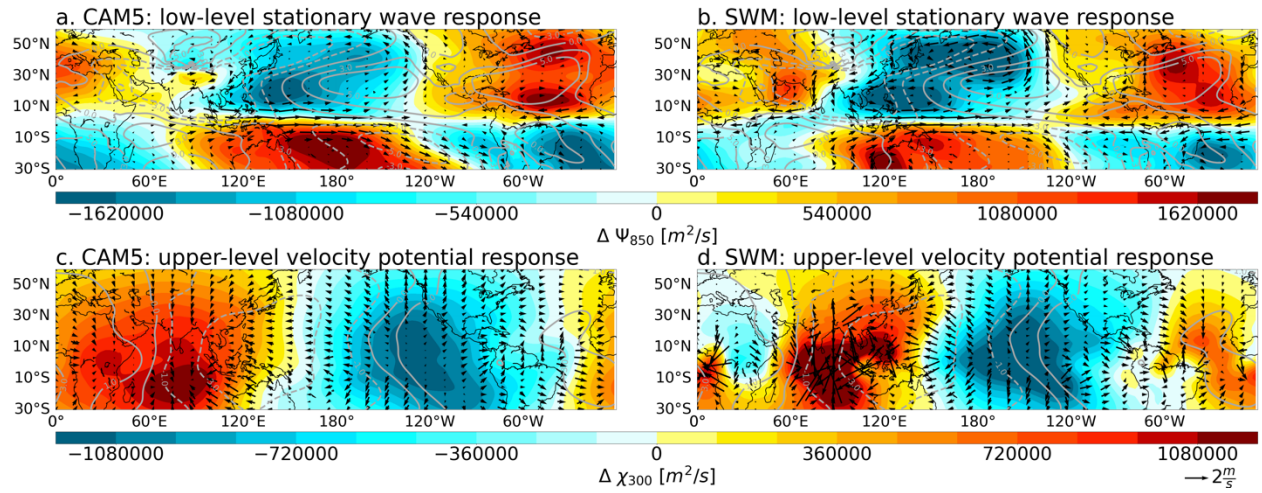


Figure S1. Circulation responses in comprehensive and stationary wave models. (a-b) 850 hPa stream function responses in amipTRP minus amipTRPadj in panel a) and in SWM-TRP minus SWM-TRPadj in panel b). (c-d) Same as panels a-b) except for 300 hPa velocity potential responses. Vectors show the corresponding rotational wind responses in panels a-b) and divergent wind responses in panels c-d).

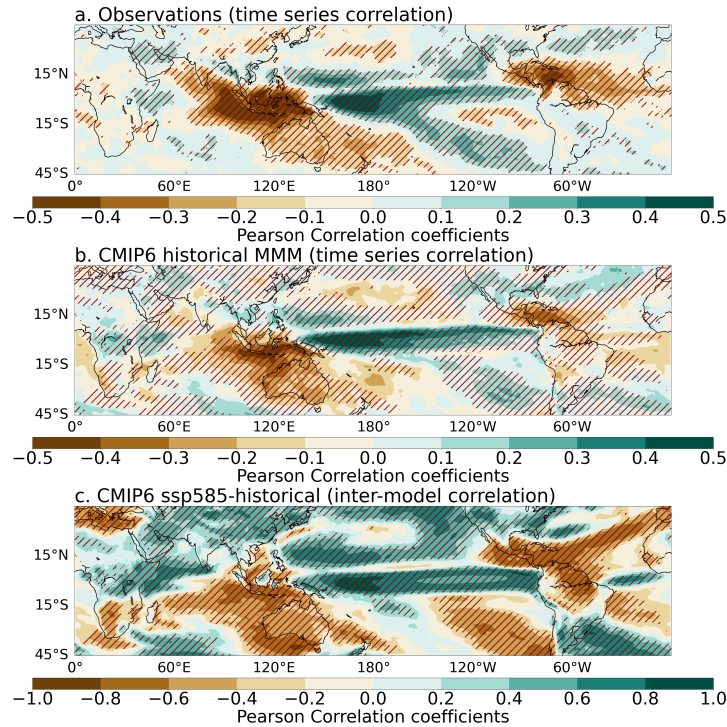


Figure S2. Connection between inter-basin precipitation differences and land precipitation. (a) Correlation between observed monthly precipitation anomalies and observed monthly I_{P-IA} anomalies. (b) CMIP6 multi-model mean correlation between monthly precipitation anomalies and monthly I_{P-IA} anomalies based the historical simulation. (c) CMIP6 inter-model correlation between future precipitation changes and future I_{P-IA} changes based the ssp585 simulation. Areas where correlation is significant at the 95% level based on a two-tailed t-test are hatched.

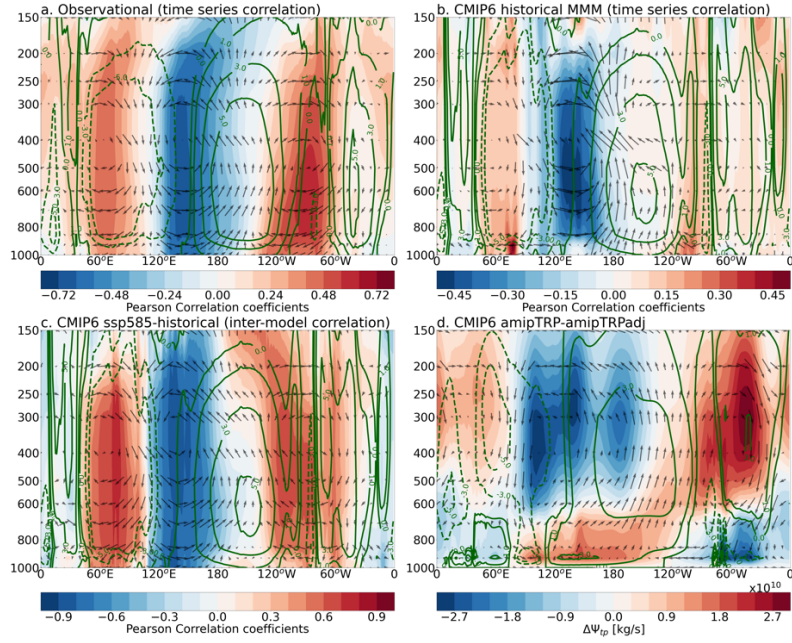


Figure S3. Vertical wind responses to inter-basin precipitation differences. (a-c) Same as Figures S2a-c), but for correlations between the tropical meridional mean mass stream function and I_{P-IA} (shading) and between the tropical meridional mean divergent wind and I_{P-IA} (vectors). (d) Responses of the tropical meridional mean mass stream function and the tropical meridional mean divergent wind from amipTRP minus amipTRPadj. Contours show the corresponding tropical meridional mean mass stream function climatologies with a contour interval of $10^{10} \text{ kg s}^{-1}$. Negative contours are dashed.

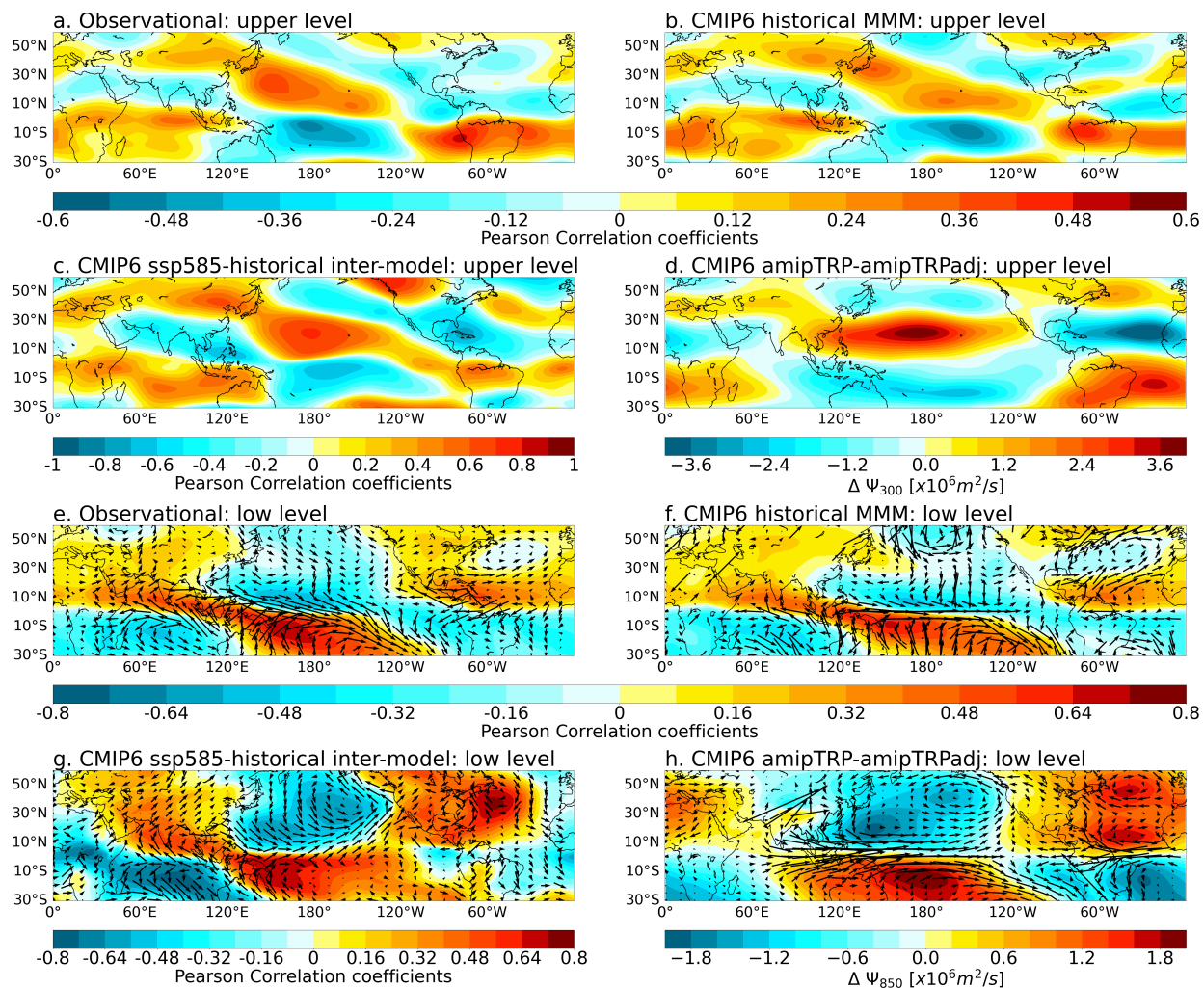


Figure S4. Circulation responses to inter-basin precipitation differences. Same as Figure S3, but for 300 hPa stream function in panels a-d) and 850 hPa stream function in panels e-h). Vectors in panels e-h) are the corresponding 850 hPa winds.

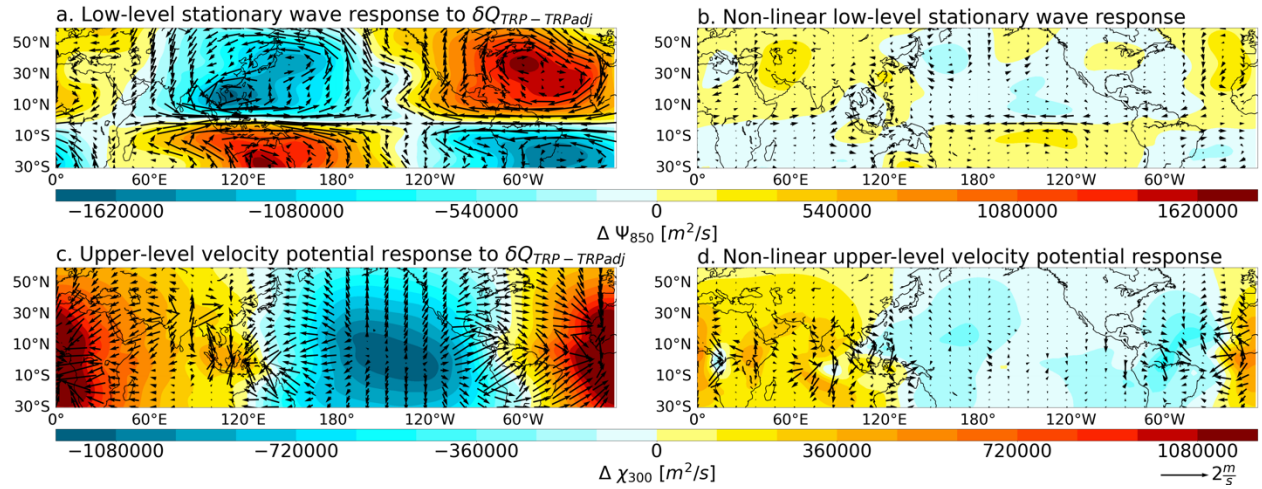


Figure S5. Linearity of circulation responses. (a) 850 hPa stream function responses in SWM-TRP-d-TRPadj. (b) differences in 850 hPa stream function responses between SWM-TRP minus SWM-TRPadj and SWM-TRP-d-TRPadj. (c-d) Same as panels a-b) except for 300 hPa velocity potential responses. Vectors show the corresponding rotational wind responses in panels a-b) and divergent wind responses in panels c-d).

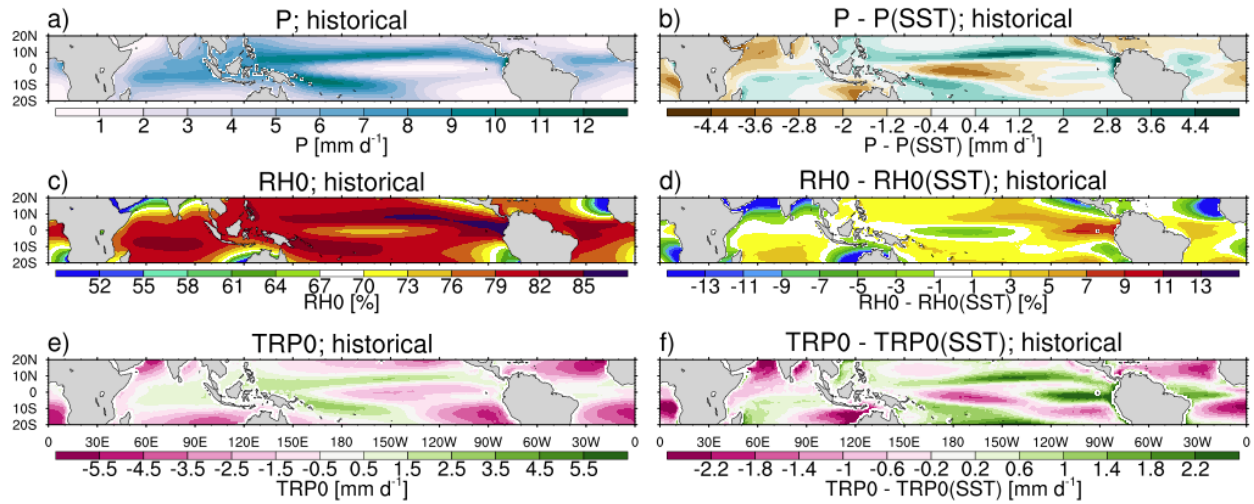


Figure S6. Relation between precipitation, relative humidity, and moisture transport. Multi-model mean present-day (a) precipitation, (c) RH0, and (e) TRP0. Multi-model mean present-day (b) precipitation, (d) RH0, and (f) TRP0 with their corresponding SST-related components removed.

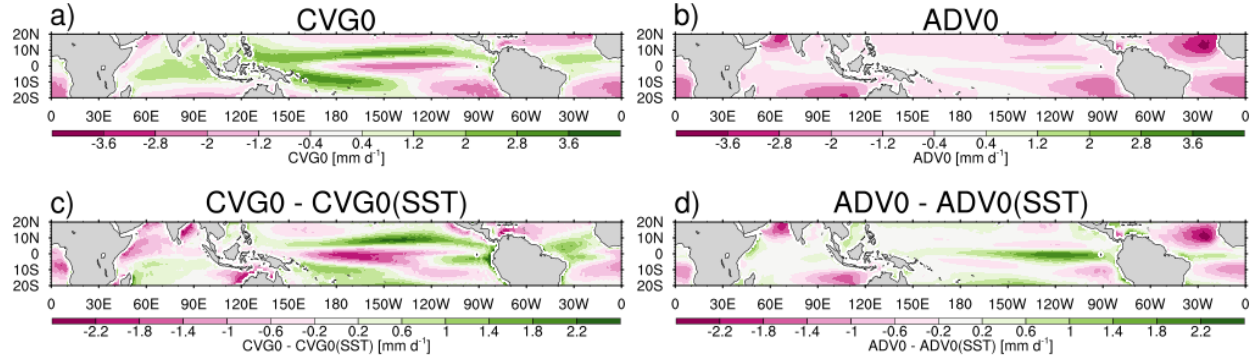


Figure S7. Moisture transport decomposition. Multi-model mean present-day (a-b) boundary layer moisture convergence (CVG0) and boundary layer moisture advection (ADV0) and (c-d) non-SST components of CVG0 and ADV0, using the historical simulation. The definition of moisture convergence and moisture advection is detailed in the Method section.

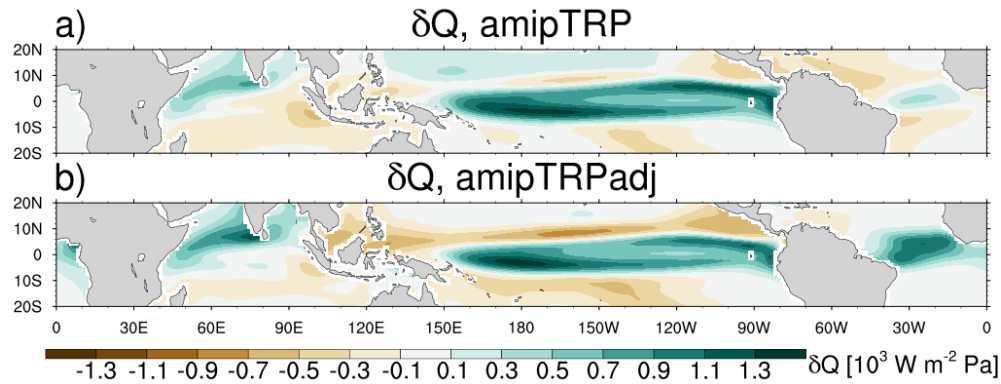


Figure S8. Diabatic heating perturbations. Simulated vertically integrated diabatic heating anomalies (δQ) from (a) amipTRP and (b) amipTRPAdj.

model name	realization used		
	historical	ssp585	amip, amip4xCO2, amip4K & amipFuture
ACCESS-CM2	rlilp1fl	rlilp1fl	X
ACCESS-ESM1-5	rlilp1fl	rlilp1fl	X
AWI-CM-1-1-MR	rlilp1fl	rlilp1fl	X
BCC-CSM2-MR*	rlilp1fl	rlilp1fl	rlilp1fl
CAMS-CSM1-0	rlilp1fl	rlilp1fl	X
CanESM5	rlilp2fl	rlilp2fl	rlilp2fl

CESM2	rlilp1f1	r4ilp1f1	rlilp1f1
CESM2-WACCM	rlilp1f1	rlilp1f1	X
CMCC-CM2-SR5	rlilp1f1	rlilp1f1	X
CMCC-ESM2	rlilp1f1	rlilp1f1	X
CNRM-CM6-1*	rlilp1f2	rlilp1f2	rlilp1f2
CNRM-CM6-1-HR	rlilp1f2	rlilp1f2	X
CNRM-ESM2-1*	rlilp1f2	rlilp1f2	X
E3SM-1-0	rlilp1f1	rlilp1f1	r2ilp1f1
E3SM-1-1	rlilp1f1	rlilp1f1	X
E3SM-1-1-ECA	rlilp1f1	rlilp1f1	X
EC-Earth3	rlilp1f1	rlilp1f1	X
EC-Earth3-CC	rlilp1f1	rlilp1f1	X
EC-Earth3-Veg	rlilp1f1	rlilp1f1	X
EC-Earth3-Veg-LR	rlilp1f1	rlilp1f1	X
FGOALS-f3-L	rlilp1f1	rlilp1f1	X
FGOALS-g3*	rlilp1f1	rlilp1f1	X
FIO-ESM-2-0	rlilp1f1	rlilp1f1	X
GFDL-CM4	rlilp1f1	rlilp1f1	rlilp1f1
GFDL-ESM4	rlilp1f1	rlilp1f1	X
GISS-E2-1-G*	rlilp1f2	rlilp1f2	X
GISS-E2-1-H*	rlilp1f2	rlilp1f2	X
GISS-E2-2-G*	rlilp3f1	rlilp3f1	X
HadGEM3-GC31-LL*	r3ilp1f3	r3ilp1f3	r5ilp1f3
HadGEM3-GC31-MM	r3ilp1f3	r3ilp1f3	X
IITM-ESM*	rlilp1f1	rlilp1f1	X
INM-CM4-8	rlilp1f1	rlilp1f1	X
INM-CM5-0	rlilp1f1	rlilp1f1	X
IPSL-CM6A-LR*	rlilp1f1	rlilp1f1	rlilp1f1
KACE-1-0-G	rlilp1f1	rlilp1f1	X
KIOST-ESM	rlilp1f1	rlilp1f1	X
MIROC6*	rlilp1f1	rlilp1f1	rlilp1f1
MIROC-ES2L*	rlilp1f2	rlilp1f2	X
MPI-ESM1-2-HR	rlilp1f1	rlilp1f1	X
MPI-ESM1-2-LR	rlilp1f1	rlilp1f1	X
MRI-ESM2-0*	rlilp1f1	rlilp1f1	rlilp1f1
NESM3	rlilp1f1	rlilp1f1	X
NorESM2-LM	rlilp1f1	rlilp1f1	X
NorESM2-MM	rlilp1f1	rlilp1f1	X
TaiESM1	rlilp1f1	rlilp1f1	rlilp1f1
UKESM1-0-LL*	rlilp1f2	rlilp1f2	X

Table S1. Models and corresponding realizations. Models that provide convective mass flux outputs are marked with *.

Supplementary Text 1: Remote Influences of Diverging Basin Hydrological Sensitivity Circulation Responses in a Stationary Wave Model

In the main text, we examine the remote impacts of diverging basin hydrological sensitivity (HS) by using the Community Atmosphere Model 5.0 (CAM5). In this section, we validate the CAM5 results by showing that the circulation responses can be reproduced in a Stationary Wave model (SWM, Method) and are consistent with the Gill-type response to tropical heating.

As shown in Figure S1, the large-scale circulation responses in amipTRP minus $\text{amipTRP}_{\text{adj}}$ and SWM-TRP minus $\text{SWM-TRP}_{\text{adj}}$ are similar. The low-level stream function responses are symmetric about the Equator. They include cyclonic circulations in the Pacific basin and anti-cyclonic circulations in the Atlantic basin, which are consistent with the classic Gill-type circulation responses to the heating in the tropical Pacific and cooling in the tropical Atlantic, respectively ¹. The upper-level velocity potential responses feature an eastward shift of the Indo-Pacific Walker cell, consistent with enhanced convection in the Pacific basin and weakened convection in the Indian and Atlantic basins.

Amazon and US Precipitation Responses in Observations and CMIP6

We hypothesize that the precipitation responses in the US and Amazon regions simulated by CAM5 (Fig. 3f) are largely driven by the Pacific-Atlantic basin HS differences, which consist of a positive (negative) diabatic heating forcing in the Pacific basin (Atlantic and Indian basins). First, we show that the simulated precipitation responses are consistent with the widely documented tropical-American teleconnections in observations. We define the inter-basin precipitation difference index ($I_{\text{P-IA}}$) as the tropical Pacific mean precipitation minus the tropical Indian and Atlantic mean precipitation. Figure S2a shows the correlation between the observed monthly precipitation anomalies and observed monthly $I_{\text{P-IA}}$ anomalies. Based on the correlation patterns in the tropical basins, a positive $I_{\text{P-IA}}$ anomaly represents the El Niño scenario, during which precipitation is repartitioned among the basins, as the Pacific basin becomes wetter and the Indian and Atlantic basins become drier ². Similar to Figure 3f, we see wetting in the US and south of Amazon and drying in Mexico, northern Amazon, Australia and much of Africa. The correlation patterns in the American continents are consistent with previous findings that the combination of a wetter Pacific and drier Atlantic basins causes precipitation to increase in the US ³ and decrease in Amazon ⁴.

The CMIP6 historical simulations capture the observed land patterns in the American continents reasonably well (Fig. S2b). The drying in the Amazon is more extensive compared to the observations. This is likely due to the drying of the South Atlantic ⁵, which is largely absent in the observations. Nevertheless, the overall consistency between observations and CMIP6 suggests that the models are able to simulate the tropical-American teleconnections. In Figure S2c, we estimate the impacts of the inter-basin HS differences in the CMIP6 multi-models through the inter-model correlation between the projected precipitation changes and the projected $I_{\text{P-IA}}$ changes. The correlation patterns over tropical oceans are similar to those in Figures S2a and S2b, but with notable differences in the Equatorial Atlantic and Northern Indian regions. Nevertheless, the drying in Southern Mexico and Amazon and the wetting in the US and south of Amazon are both significantly correlated with the inter-basin HS differences. Note that the oceanic precipitation patterns in Figures S2a-c are somewhat different, so we do not expect the corresponding land precipitation patterns to all exactly match that in Figure 3f. The above results are all supportive of the precipitation responses in the American continents simulated by CAM5.

Mechanisms of Precipitation Responses in the Amazon and US

Mechanisms of the observed Amazon and US precipitation responses to Pacific wetting and Atlantic drying have been long established and we found these mechanisms to apply to the CAM5 responses as well.

The drying in the Amazon region is mainly associated with changes in the Walker cell ^{4,6,7}. As shown in Figure S3, the Pacific wetting and Atlantic drying correspond to an eastward shift of the Walker cell, as the ascending branch moves from the Maritime Continents towards Central Pacific and the descending branch moves from Eastern Pacific to South America and the Atlantic basin. In addition, the Pacific wetting drives surface easterlies to its east as a Kelvin wave response and the Indian and Atlantic drying drives surface westerlies to its west as a Rossby wave response ¹. Such responses in surface winds create a surface divergence in the Amazon region, which further contributes to the decrease in precipitation.

As shown in Figures S4e-h, the wetting in the US region is mainly driven by the poleward low-level winds which transports warm, moist air from the tropical Eastern Pacific to the coast of California ^{8,9} and from the Caribbean Sea to Southeast US ³. These poleward winds are associated with the western flank of the cyclonic circulation in the tropical North Pacific and the eastern flank of the anticyclonic circulation in the US and the tropical North Atlantic region. The former is the Gill response ¹ to the positive diabatic heating forcing in the Pacific basin. The latter is a combination of the Gill response to the negative diabatic heating forcing in the Atlantic basin and the surface manifestation of the upper-level Rossby wave train associated with the Pacific North American pattern that consists of anticyclonic circulations in the tropical North Pacific and over Canada and cyclonic circulations in the extratropical North Pacific and over the US (Figs. S4a-d).

Linearity of Responses to Diabatic Heating Perturbations

Because we use $\text{amipTRP} - \text{amipTRP}_{\text{adj}}$ to represent the impacts of inter-basin differences in $\text{HS}_{\text{local}\%}$, we assume that the responses to the diabatic heating perturbations are linear. We have shown that the circulation responses in $\text{amipTRP} - \text{amipTRP}_{\text{adj}}$ conform to the patterns derived by Gill's linear model ¹. In this section, we further examine the linearity assumption with the SWM, which is a nonlinear model. By comparing $\text{SWM-TRP} - \text{SWM-TRP}_{\text{adj}}$ (Figs. S1b, d) with $\text{SWM-TRP-d-TRP}_{\text{adj}}$ (Method, Figs. S5a, c), we find overall similar circulation patterns between the two. The effects of nonlinearity are far smaller than total responses (Figs. S5b, d).

Supplementary Text 2: Inter-basin Differences in Boundary-layer Relative Humidity

To understand why boundary-layer relative humidity (RH0) differs among basins at given SSTs, we begin by examining the spatial distribution of RH0 (Fig. S6c). To first order, RH0 is determined by the underlying SSTs, as warm regions are often associated with abundance of saturated air plumes and lack of dry subsidence, which leads to high RH0¹⁰. On the other hand, humidity is also affected by horizontal moisture transport, which varies substantially across tropical oceans¹¹.

To understand why RH0 differs among basins at given SSTs, we remove the part of RH0 that is directly related to the amplitude of SSTs. Specifically, we calculate the SST component of RH0 as the average of RH0 over tropical oceans for 0.1 °C SST_{rel} bins. This is essentially the same as transforming RH0 from X-Y space to SST_{rel} space (see Figs. 4e, f) but for the entire tropical oceans instead of individual basins. Note that the SST component only accounts for the influence of SST amplitude on RH0 but does not account for the effect of SST gradients. We then obtain the non-SST component of RH0 by subtracting the SST-component of RH0 from the total RH0. The non-SST component shows spatial variations in RH0 unrelated to the amplitude of SSTs (Fig. S6d) and explains why RH0 differs among basins in SST_{rel} space (Figs. 4e, f). The relatively high RH0 in the Pacific basin is mainly associated with two features – the lack of low RH0 in the subtropical area compared to the other basins and the presence of high RH0 in the northeastern Equatorial Pacific regions. By examining boundary-layer moisture transport (TRP0, Method) and its non-SST component, we find that spatial variations in RH0 at given SSTs are primarily driven by moisture transport (Figs. S6e, f). The high degree of spatial resemblance between precipitation, RH0, and TRP0 is present in observations and reanalyses as well (not shown).

By separating TRP0 into a part driven by moisture advection and a part driven by moisture convergence (Method, Fig. S7), we find that the negative non-SST component of RH0 in the subtropical regions of the Indian and Atlantic basins is primarily associated with the advection of dry air (Fig. S7d), as these regions are located downwind of arid lands¹². On the other hand, the positive non-SST component of RH0 in the northeastern Equatorial Pacific region is associated with the strong moisture convergence (Fig. S7c), which results from local pressure gradients driven by the underlying SST fronts¹³.

In summary, less dry advection from land and greater moisture convergence from sharp SST gradients yield a more positive TRP0 in the Pacific basin. This explains the higher RH0 in the Pacific basin, which causes higher MSE0 and ultimately, larger precipitation.

References

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