Optimal geometric characterization of forced zonal mean tropical precipitation changes.

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Abstract The zonal and annual mean tropical precipitation response to paleoclimate and anthropogenic forcing scenarios ranging from the Last Glacial Maximum (LGM), CO₂ quadrupling (4XCO₂), mid-Holocene, North Atlantic freshwater hosing and volcanic forcing is analyzed in an ensemble of global climate models. Zonally averaged tropical precipitation changes are characterized in terms of three geometric manipulations of the climatological precipitation

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(hereafter, modes): meridional shifts, intensifications, and meridional contractions. We employ an optimization procedure that quantifies the magnitude and robustness (across different models) of changes in each mode in response to each forcing type. Additionally, the fraction of precipitation changes that are explained by the modes (in isolation and combined) is quantified. Shifts are generally less than 1° latitude in magnitude and explain a small fraction ($<10\%$) of tropical precipitation changes. Contractions and intensifications are strongly anti-correlated across all simulations with a robust intensification and contraction of precipitation under global warming and a robust reduction and expansion under global cooling during the Last Glacial Maximum. The near constant scaling between contractions and intensifications across all simulations is used to define a joint contraction/intensification (*CI*) mode of tropical precipitation. The *CI* mode explains nearly 50% of the precipitation change under $4XCO_2$ and LGM forcing by optimizing a single parameter. These results suggest the shifting mode that has been extensively used to interpret paleo-rainfall reconstructions is of limited use for characterizing forced zonal mean precipitation changes and advocates for a reinterpretation of past precipitation changes to account for the *CI* mode.

Keywords Tropical Precipitation · ITCZ · Paleoclimate · Tropical Contraction

1 Declarations

1.1 Funding

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1.2 Conflicts of interest/Competing interests

None.

1.3 Availability of data and material

The model precipitation data used in this study were uploaded to Figshare and published with the following doi (10.6084/m9.figshare.12284282). All CMIP5/PMIP3 and CMIP3/PMIP2 climate model data were downloaded from the Earth System Grid Federation (ESGF) node hosted by Lawrence Livermore National Laboratory data repositories (<https://esgf-node.llnl.gov/>).

1.4 Code availability

Not applicable.

1.5 Authors contributions

AD formulated the optimization procedure, analyzed the model simulations and wrote the manuscript. ARA designed the freshwater hosing experiments,

compiled model simulations and edited the manuscript. DSB provided feedback on model design, edited manuscript and provided feedback on which colors (shades of pink) readers find generally offensive in graphics.

2 Introduction

The concept of a meridional shift of the intertropical convergence zone (ITCZ) is ubiquitous in both the paleoclimate and climate dynamics literature for describing spatial patterns of forced tropical precipitation changes. Here, we define an ITCZ shift as a meridional translation of the climatological tropical precipitation pattern. In paleoclimate studies, concurrent increases and decreases of precipitation in different latitude bands within the tropics have commonly been interpreted as a meridional translation (shift) of the ITCZ including: (i) southward ITCZ shifts in response to Heinrich events (*Jacobel et al.*, 2016; *Wang et al.*, 2001) characterized by North Atlantic iceberg discharge; (ii) southward ITCZ shifts during the Little Ice Age (*Pahnke et al.*, 2007; *Haug et al.*, 2001; *Sachs et al.*, 2009) when proxy evidence suggests the Northern Hemisphere was cold relative to the rest of the Holocene; (iii) northward ITCZ shifts during the early Holocene (*Haug et al.*, 2001; *Bird et al.*, 2011; *Wang et al.*, 2001; *Sachs et al.*, 2018) when precessional phasing intensified boreal summer insolation; and (iv) southward ITCZ shifts during the Last Glacial Maximum (*Arbuszewski et al.*, 2013; *Koutavas and Lynch-Stieglitz*, 2004) when large ice sheets occupied the Northern Hemisphere continents. A meridional translation of a strongly peaked climatological precipitation dis-

tribution would result in a dipole of precipitation anomalies. Thus, the contrasting precipitation changes across different latitude bands discerned from compilations of paleoclimate records are qualitatively consistent with an ITCZ shift (*Arbuszewski et al.*, 2013; *Sachs et al.*, 2009).

In idealized dynamical studies (i.e. aquaplanet simulations) forced by hemispherically asymmetric heating/cooling, the ITCZ very clearly shifts towards the hemisphere in which the atmosphere is heated more strongly. This ITCZ response is commonly understood in terms of the Hadley cell shifting in order to redistribute the excess energy in the warmed hemisphere to the opposite hemisphere (*Kang et al.*, 2008; *Wei and Bordoni*, 2018; *Frierson and Hwang*, 2012). The same zonal mean ITCZ shift in response to hemispherically asymmetric forcing is also seen in comprehensive coupled climate models (*Chiang and Bitz*, 2005; *Schneider et al.*, 2014; *Donohoe et al.*, 2013). However, the spatial pattern of forced precipitation changes in models with zonal asymmetries such as geography, ocean circulations and a seasonal cycle is more complicated than a simple shift of the climatological precipitation pattern. ITCZ shifts in comprehensive models are diagnosed in a myriad of ways, including changes in the “precipitation centroid” (P_{CENT} — akin to a gravitational center of mass) (*Frierson and Hwang*, 2012), the hemispheric contrast of tropical precipitation (*Hwang et al.*, 2013), the precipitation power weighted latitude or “moment” (*Adam et al.*, 2016) or by simply noting the spatial map of precipitation change is characterized by a meridional dipole. These various ITCZ shift metrics have different strengths and weaknesses, but in nearly all cases,

little effort has been made to quantify how much of the forced precipitation changes can be explained by an ITCZ shift.

Consider the following example of how a tropical precipitation change that does not involve a meridional translation of precipitation might be diagnosed as an “ITCZ shift” starting from a climatological tropical precipitation distribution is bi-modal with peaks north and south of the equator – as is common in many climate model simulations (e.g. the black line in Fig. 1A). Assume that, in response to external forcing, the precipitation peak in the NH is intensified with no other changes in the distribution. This precipitation change would be diagnosed as “northward shift” in the common metrics used in the literature (e.g. P_{CENT} , precipitation weighted moments, hemispheric contrast of precipitation). However, a meridional translation of the climatological precipitation is a poor characterization of the precipitation change since the locations of the tropical precipitation peaks have not moved.

In this manuscript, we quantify how well a meridional shift in the climatological zonal mean precipitation pattern characterizes simulated precipitation changes (ΔP) under a myriad of climate forcings in the following way. The climatological zonal mean tropical precipitation distribution is shifted north and south in search of the resultant pattern that best matches the simulated precipitation in response to forcing; the fraction of ΔP explained by the shift is quantified by how well the optimally shifted distribution matches the forced response. This procedure also provides a new metric for quantifying the magnitude of ITCZ shifts (optimized value of the translation in degrees latitude)

that can be compared against other commonly used metrics to address how far the ITCZ shifts in model simulations. Additionally, we ask if there are other geometric manipulations of the climatological precipitation (hereafter, modes) that explain a larger fraction of simulated tropical precipitation changes than the shifting mode. Specifically we analyze two additional modes: i) intensification of the precipitation distribution with no change in shape and ii) meridional contraction/expansion of the precipitation distribution about the equator. These three modes — shifting, intensifying and contracting — are optimally fit, in conjunction, to best match the forced changes in precipitation.

We consider forced precipitation changes in large ensembles of climate models across a myriad of different climate forcings including: i) CO₂ forcing; ii) Last Glacial Maximum orbital parameters and boundary conditions; iii) mid-Holocene orbital parameters and greenhouse gas concentrations; iv) freshwater hosing of the North Atlantic and v) volcanic forcing. We limit our analysis to annual and zonal mean precipitation changes (hereafter ΔP) in the current manuscript, to focus on characterizing the modes of tropical precipitation changes on the broadest spatial-temporal scale while acknowledging that zonally-homogeneous and seasonal changes of tropical precipitation are robust and comparable in magnitude to zonal, annual mean changes (*Atwood et al.*, 2020; *Donohoe et al.*, 2019; *Haug et al.*, 2001; *Yan et al.*, 2015; *Braconnot et al.*, 2012; *Liu et al.*, 2018).

This manuscript is organized as follows. The model simulations that we analyze and the methodology for optimally characterizing the tropical precipita-

tion changes are described in Section 3. The resulting optimal characterization of ΔP in terms of shifts, intensifications and, contractions and the fraction of precipitation changes described by each mode across the different prescribed forcing scenarios are described in Section 4. A summary and discussion follows.

3 Model simulations analyzed and methodology

Here we describe the suite of coupled climate model simulations and forcing scenarios for which ΔP is analyzed and the methodology for optimally characterizing ΔP in terms of shifting, intensifying and contracting modes.

3.1 Model simulations analyzed

We make use of model output for idealized anthropogenic (abrupt CO₂ quadrupling) and paleoclimate (LGM, mid-Holocene) forcing scenarios run across an ensemble of different climate models as part of the Climate Model Inter-comparison Project (*Taylor et al.*, 2012) and Paleoclimate Model Inter-comparison Project (*Braconnot et al.*, 2007a, 2012). Additional volcanic and freshwater forcing simulations are analyzed in a subset of coupled climate models. *Atwood et al.* (2020) analyzed the zonal structure of tropical precipitation changes in the same suite of model simulations and additional details of the simulations are provided there. 95 sets of forced simulations and companion control simulations are analyzed.

CO₂ quadrupling: 22 simulations from the abrupt CO₂ quadrupling experiment (hereafter 4XCO₂) of CMIP5 (*Taylor et al.*, 2012) are analyzed

in which atmospheric CO₂ concentrations are instantly quadrupled relative to pre-industrial (PI) levels and held fixed thereafter. ΔP is evaluated from the difference between the climatology in years 100-150 after quadrupling in the 4XCO₂ simulations and that in the PI control simulation in the same model.

Last Glacial Maximum – LGM: 13 total LGM simulations are analyzed with 7 from PMIP2 (*Braconnot et al., 2007a*) and 6 from PMIP3 (*Braconnot et al., 2012*). LGM simulations are forced by solar insolation from 21,000 years before present, greenhouse gas (CO₂, CH₄, N₂O) and aerosol concentrations are set based on ice core data (CO₂ is set to 185 ppm) coastlines are set to be consistent with a 120m decrease in sea level, and the land ice topography is prescribed as specified in *Kageyama et al. (2018)* which includes the expansive Laurentide ice sheet over North America. ΔP is evaluated from the LGM climatology calculated from years 31-200 after a spin up period.

Mid Holocene: 12 PMIP3 (*Braconnot et al., 2012*) mid-Holocene simulations are analyzed. Simulations are forced by orbital parameters representative of 6,000 years before present from *Berger (1978)*. Most significantly, due to orbital precession, the seasonal cycle of insolation in the Northern Hemisphere was amplified 6,000 years ago while the duration of summer was reduced. We note that in the annual average the sum of all mid-Holocene forcing agents are symmetric about the equator.

Freshwater Hosing: 18 total simulations of North Atlantic freshwater forcing (aka 'hosing') are analyzed – 16 from the Community Climate System Model version 4 (CCSM4) as described in *Atwood et al. (2020)* and 1 each with

Max Planck Institute of Meteorology OM1 model and the United Kingdom Hadley Center CM3 model as described in *Timmermann et al. (2007)*. In each simulation either 1 Sv or 0.1 Sv of freshwater forcing is imposed across the surface of the North Atlantic Ocean (from 50°N-70°N) for 100 years. ΔP is calculated from the climatology averaged over the last 80 years of simulation (it takes approximately 20 years for the Atlantic Meridional Overturning circulation to fully respond to the freshwater forcing). Four ensemble members were performed with the default configuration of CCSM4 with 1 Sv meltwater forcing. The same freshwater forcing was also applied to two bias-corrected versions of the model. The bias corrections were: (1) a modification to the topography of Central America (which reduced the biases in the eastern Pacific climatology associated with the poor resolution of Central American topography; *Baldwin et al., 2021*) and (2) surface heat flux modifications with a cyclostationary seasonal cycle throughout the tropical oceans (30°S to 30°N) to further reduce the bias in the climatological seasonal cycle in SST. In one configuration of the model with three ensemble members, only the Central American topography was modified with no changes in the surface heat fluxes. In a second configuration of the model with four ensemble members, both the topography and surface heat flux corrections were prescribed. A final set of four ensemble members was performed with 0.1 Sv meltwater forcing with both topography and surface heat flux corrections prescribed (see *Atwood et al., 2020*, for more details).

Volcanic Forcing: The climate response to volcanic forcing is analyzed separately for volcanic events localized in the NH extratropics, SH extratropics and tropics. Volcanic events from two different types of simulations are analyzed: i) volcanic forcing only experiments and ii) last millennium simulations with all climate forcings.

The volcanic forcing only simulations consist of 5 CESM4 Last Millennium Ensemble (LME) volcanic-only simulations (with separate composite volcanic responses constructed for eruptions located in the NH, SH and tropics – making 15 CESM4 composites) and an ensemble of simulations with the Norwegian Earth System Model Version 1-M (NorESM) mimicking a high-latitude Northern Hemisphere (NH) summer eruption (the Laki eruption in Iceland; *Pausata et al.*, 2015a,b). In the NorESM simulations, 100 Tg of SO₂ and dust was added to the upper atmosphere over a four month period to simulate the Laki eruption. Eruptions were prescribed at a different starting year of a transient historical simulation (1850-2005) for each of the 48 ensemble members. 3 composites each consisting of 16 ensemble members are formed to be consistent with the number of eruptions in the last millennium composites (discussed below). The NorESM volcanic forcing runs are compared against control runs in the same model that were branched from the same transient historical simulation but with no prescribed volcanic forcing.

The response to volcanic forcing is also assessed from selected PMIP3 all-forcing last millennium (LM) transient simulations (CCSM4 and GISS Model E Ensemble Members 122, 125, and 128). In the last millennium simulations,

ΔP forced by volcanic eruptions is diagnosed from the composite response in the years after a volcanic event minus that in the closest 5 years preceding the event with minimal volcanic activity. Specifically, for each eruption, the monthly precipitation response was calculated from months with AOD > 0.02 up to 2 years before and after the peak of the eruption and compared to the control climatology, which was taken to be the climatology of the closest 5-year period prior to the onset of the event with minimal volcanic activity (defined as global mean AOD summed over the 5-year period < 0.01). Tropical volcanic events are defined as volcanic events with peak globally-averaged AOD > 0.1 , and extratropical volcanic events are further defined by hemispherically-averaged AOD at least 25% greater in one hemisphere. This procedure results in composites that consist of 9-20 events in each region and simulation (See *Atwood et al.*, 2020, for more details)..

3.2 Methodology for defining optimal precipitation shifts, intensifications and, contractions

We characterize the forced ΔP in terms of shifts, intensifications and contractions of the unforced (control) climatological tropical precipitation by optimally geometric manipulating the control precipitation to best match the forced precipitation. We begin by evaluating the optimal shift acting in isolation. To search for the optimal shift, the unperturbed (pre-industrial) zonal mean precipitation, $P_{PI}(\theta)$, is shifted in latitude, θ , both northward and southward by S degrees latitude in increments of 0.02° up to 5° . Mathematically:

$$P_{SHIFT}(\theta) = P_{PI}(\theta - S). \quad (1)$$

For each value of S in the search algorithm, the fraction of ΔP explained by a shift of S ($FRAC_{SHIFT}$) is defined as:

$$FRAC_{SHIFT} = 1 - \frac{\sqrt{\langle (P_{FORCED} - P_{SHIFT})^2 \rangle}}{\sqrt{\langle \Delta P^2 \rangle}} \quad (2)$$

where angled brackets, $\langle \rangle$, denote spatial averages over the tropics (defined as equatorward of 20°). The fraction of ΔP explained by the shift is negatively proportional to the root mean square (RMS) difference between the shifted precipitation (P_{SHIFT}) and the target precipitation (P_{FORCED}) normalized by the root mean square of ΔP . If P_{SHIFT} perfectly matches P_{FORCED} , the RMS difference is zero and the fraction explained is 1. Whereas for a shift of zero, $P_{FORCED} - P_{SHIFT,0} = \Delta P$ so the numerator and denominator are equal and the $FRAC_{SHIFT} = 0$. The optimal shift is chosen such that $FRAC_{SHIFT}$ is maximized or, equivalently, the RMS error between the shifted precipitation and the target precipitation (P_{FORCED}) is minimized.

To demonstrate the optimization algorithm, consider the precipitation response to freshwater forcing in the CCSM4 model shown by the green line in Fig. 1B which has enhanced tropical precipitation in the SH tropics relative to that in the unforced simulation (P_{PI} , black line). The goal is to meridionally translate P_{PI} (black line) to optimally fit P_{FORCED} (green line). The optimal shift is determined to be -1.7° (with negative indicating southward) and the optimally shifted precipitation is shown by the blue curve. $FRAC_{SHIFT} = 0.251$

which indicates that 25.1% of ΔP is explained by the shift. By construction, the optimally shifted precipitation has the same shape as P_{PI} but is meridionally translated; the significant mismatch between P_{SHIFT} and P_{FORCED} is associated with changes in the amplitudes of the peaks. The optimally shifted precipitation matches P_{FORCED} significantly better than P_{PI} but it is far from a perfect match. We note the optimal shift is evaluated to within 0.02° latitude which is a finer resolution than the model output. These sub-grid scale shifts impact precipitation changes on the coarser model grid.

Next we optimally shift and intensify P_{PI} in conjunction to best match P_{FORCED} to ask: what additional fraction of ΔP can be explained by introducing an optimal intensification of the climatological precipitation (relative to the optimal shift only)? Mathematically, the combined shift and intensification of P_{PI} is

$$P_{SHIFT,INT}(\theta) = I \cdot P_{PI}(\theta - S), \quad (3)$$

where I is the (dimensionless) intensification scalar. We calculate $P_{SHIFT,INT}$ for values of I between 0.8 and 1.2 in increments of 0.01 and over the S values previously considered in the shift only case. The optimal values of S and I are defined as those for which the RMS difference between $P_{SHIFT,INT}$ and P_{FORCED} is minimized — noting that the optimal value of S for the combined shift and intensification need not equal that for the optimal shift only. The additional fraction of ΔP explained by the intensification, $FRAC_{INT}$, is

defined as the improvement of RMS error in the fit to P_{FORCED} relative to that of the optimally shifted only precipitation, normalized by RMS ΔP :

$$FRAC_{INT} = \frac{\sqrt{\langle (P_{FORCED} - P_{SHIFT})^2 \rangle}}{\sqrt{\langle \Delta P^2 \rangle}} - \frac{\sqrt{\langle (P_{FORCED} - P_{SHIFT,INT})^2 \rangle}}{\sqrt{\langle \Delta P^2 \rangle}}. \quad (4)$$

Returning to our previous example of the precipitation response in CCSM4 to freshwater hosing, the improved fraction of ΔP explained due to introducing an optimal intensification can be visualized in Fig. 1B by comparing the mismatch between the target precipitation (P_{FORCED} , green line) and the shifted and intensified precipitation ($P_{SHIFT,INT}$, red) versus that for the shifted precipitation (P_{SHIFT} , blue). A clearer example of contribution by the intensification mode is given by the simulated precipitation response to $4XCO_2$ in the FGOALS model (Fig. 1A). In this case, the shift alone (optimal value of 0.5° northward) describes a very small fraction of ΔP (1.7%) whereas the combined shift and intensification (optimized $I = 1.16$ intensification) explain 38.2% of ΔP . The difference in fraction of ΔP explained between these two optimized fits (37.5%) is defined as $FRAC_{INT}$.

Next, we introduce a meridional contraction of P_{PI} in conjunction with the shifting and intensifying modes using the form

$$P_{SHIFT,INT,CON}(\theta) = I \cdot P_{PI} \left(\frac{1}{C} \cdot (\theta - S) \right), \quad (5)$$

where C is the contraction scalar with $C > 1$ corresponding to an expansion and $C < 1$ corresponding to a contraction. We search over values of C between

0.8 and 1.2 in increments of 0.01 and over the S and I values previously considered. The optimal values of S , I and, C are defined as those for which the RMS difference between $P_{SHIFT,INT,CON}$ and P_{FORCED} is minimized — noting that the optimal value of S and I for the optimal $P_{SHIFT,INT,CON}$ need not equal those defined for $P_{SHIFT,INT}$ and/or P_{SHIFT} . The additional fraction of ΔP explained by the contraction, $FRAC_{CON}$, is defined as the improvement of RMS error in the fit to P_{FORCED} relative to that of the optimally shifted and intensified precipitation ($P_{SHIFT,INT}$) normalized by the RMS ΔP :

$$FRAC_{CON} = \frac{\sqrt{\langle (P_{FORCED} - P_{SHIFT,INT})^2 \rangle}}{\sqrt{\langle \Delta P^2 \rangle}} - \frac{\sqrt{\langle (P_{FORCED} - P_{SHIFT,INT,CON})^2 \rangle}}{\sqrt{\langle \Delta P^2 \rangle}} \quad (6)$$

The improved fraction of ΔP explained due to introducing an optimal contraction can be visualized in Fig. 1A by the mismatch between P_{FORCED} (green) and $P_{SHIFT,INT,CON}$ (orange) as compared to that of $P_{SHIFT,INT}$ (red). In the example of the simulated precipitation response to $4XCO_2$ in the FGOALS model (Fig. 1A), the intensity of the bi-modal peaks (near $7^\circ N$ and $7^\circ S$) in the optimized $P_{SHIFT,INT}$ (before the contraction mode is introduced) is too small whereas the intensity of the precipitation on the poleward flanks of the peaks is too large and, thus, further increasing the I scalar in $P_{SHIFT,INT}$ to match the peaks height increases the RMS mismatch with P_{FORCED} . However, introducing a contraction in conjunction with a shift

and intensification of the precipitation distribution allows the optimal solution to match both the increased precipitation in the peaks of the distribution and the nearly unchanged precipitation on the flanks. The combination of all three optimized modes ($P_{SHIFT,INT,CON}$) explains an impressive 75.4% of ΔP . This improved fit relative to $P_{SHIFT,INT}$, which only explains 39.2% of ΔP , is attributed to the contraction mode ($FRAC_{CON} = 36.2\% = 75.4\% - 39.2\%$).

We have chosen to apply the precipitation modes in the order: (1) shift, (2) intensification, (3) contraction. This methodology answers the following sequence of questions: i) What fraction of ΔP is described by an optimal shift alone? ii) How much additional ΔP can be described by adding an optimal intensification in conjunction with the shift? and iii) How much additional ΔP can be described by introducing an optimal contraction in addition to the shift and intensification? Because the three precipitation modes are not generally spatially orthogonal, the order of applying the modes does impact the resultant $FRAC_{SHIFT}$, $FRAC_{INT}$ and $FRAC_{CON}$. In practice, the shifting mode is nearly spatially orthogonal to the other modes whereas the contraction and intensification are not. This characteristic can be understood as follows; since precipitation is larger in the deep tropics (i.e. near the equator) than at the edges of the tropical domain, an expansion of P_{PI} ($C > 1$) results in a domain wide intensification and, thus, projects onto the intensifying mode. Qualitatively similar results for $FRAC_{SHIFT}$, $FRAC_{INT}$ and, $FRAC_{CON}$ are found for the five alternative permutations of the order of modes aside from one ex-

ception: the value of C found for optimizing ΔP is significantly impacted by whether contraction is introduced before or after the intensification mode due to the previously mentioned increase in tropical domain average precipitation that results from meridional expansion alone. For this reason, we report the modal parameters (S , I and, C) that result from optimizing all three modes simultaneously. We report $\text{FRAC}_{\text{SHIFT}}$, FRAC_{INT} and, FRAC_{INT} that result from the specific order of modes outlined above.

4 Results

We begin by characterizing the ensemble mean and spread of ΔP under each forcing scenario in terms of optimal shifts, intensifications and contractions of unperturbed precipitation (P_{PI}). In Section 4.2 we then generalize across experiment type to assess the relative importance of shifts, intensifications and contraction for describing tropical precipitation changes and place our results concerning the shifts in the context of the existing literature on ITCZ shifts in Section 4.3.

4.1 Characterization of tropical precipitation changes in response to different forcings

4.1.1 Last Glacial Maximum and $4XCO_2$

The tropical precipitation changes from the cold LGM climate (solid blue line in Fig. 2A) to the pre-industrial (black line) and from the pre-industrial

to the warm 4XCO₂ climate (solid red line) are best characterized as an intensification and meridional contraction of the precipitation distribution as the planet warms. The precipitation contraction with warming is visually evident from the equatorward migration of the bi-modal precipitation peaks in warmer climates, especially the SH peak under 4XCO₂ forcing. The optimally shifted, intensified and contracted fit ($P_{SHIFT,INT,CON}$) to P_{FORCED} is shown by the dashed lines in Fig. 2A, which captures a large fraction of ΔP ; $P_{SHIFT,INT,CON}$ explains 62% and 61% of the ensemble mean ΔP for the LGM and 4XCO₂ simulations. The optimal modal fit of the LGM precipitation ($P_{SHIFT,INT,CON}$) features a robust decrease in the intensity of tropical precipitation across models ($I = 0.91 \pm 0.02$) and an expansion of the precipitation distribution ($C = 1.04 \pm 0.02$) where the stated range is two standard deviations of the ensemble mean, roughly corresponding to the 95% confidence interval (values are shown in the table inset of Fig. 2A). In contrast, in response to 4XCO₂ forcing there is robust intensification of tropical precipitation ($I = 1.13 \pm 0.02$) and a contraction in the precipitation distribution ($C = 0.89 \pm 0.02$). There is also a modest but robust southward shift in precipitation under LGM forcing ($S = -0.4^\circ \pm 0.2^\circ$), whereas the shift under 4XCO₂ is not robust — a southward shift is found in 9 models and northward shift is found in 13 models, in agreement with the analysis of *Donohoe and Voigt* (2015).

The optimal shift under LGM forcing is southward in 11 of the 13 simulations (abscissa in Fig. 3A) with an ensemble mean value of -0.4° and is $< 1^\circ$ in the majority of models. The magnitude of these shifts is almost an order

of magnitude less than those inferred at regional scales from paleoclimate reconstructions such as the 7° southward shift of the Atlantic ITCZ suggested by *Arbuszewski et al.* (2013). Similarly, the magnitude of the optimal shift in response to 4XCO_2 is less than 1° in all ensemble members (red dots in Fig. 3A). In addition to the small magnitude of the optimal shift, the shifting mode explains a small fraction of ΔP under both LGM and 4XCO_2 forcing; the ensemble average $\text{FRAC}_{\text{SHIFT}}$ is 4.1% and 2.4% for the LGM and 4XCO_2 simulations and exceeds 10% in only two of the LGM simulations and one of the 4XCO_2 simulations (Fig. 4A).

In addition to the robust ensemble mean intensification and contraction of tropical precipitation under global warming, the intensification and contraction are highly (negatively) correlated across model simulations, with correlation coefficients of -0.95 and -0.52 for the LGM and 4XCO_2 simulations, respectively (Fig. 3B). This result suggests that the degree of contraction and the degree of intensification of the tropical precipitation distribution in response to forcing go hand and hand. We return to this point in Section 4.3. The intensification and contraction modes each explain a significantly greater fraction of ΔP than the shift mode in response to LGM and 4XCO_2 forcings (Fig. 4A). The intensification mode explains 47.5% and 29.7% of ΔP in the LGM and 4XCO_2 simulations, respectively, with values exceeding 10% in all of the LGM simulations and 19 of 22 of the 4XCO_2 simulations (Fig. 4A). The contraction mode explains 10.3% and 29.1% of ΔP in the LGM and 4XCO_2 simulations, respectively, with values exceeding 10% in 7 of 13 of the LGM

simulations and 20 of 22 of the 4XCO₂ simulations (Fig. 4A). These results collectively demonstrate that contractions and intensifications of tropical precipitation respond robustly to CO₂ and LGM forcing and better characterize the changes in tropical precipitation changes than does meridional shift.

4.1.2 Freshwater hosing

The ensemble mean tropical precipitation under 1 Sv of freshwater hosing of the North Atlantic (solid orange in Fig. 2B) features a reduction in the peak amplitude in the NH and an amplification of the precipitation maximum in the SH relative to precipitation in the PI simulation (black solid line). ΔP in response to freshwater hosing is best characterized by the shifting mode (ensemble mean $\text{FRAC}_{\text{SHIFT}}$ of 20.3% — Fig. 4A) with a robust ensemble mean southward shift of $-1.4^\circ \pm 0.3^\circ$ (Fig. 2B). ΔP in response to 0.1 Sv of freshwater forcing (4 CCSM4 simulations— not shown) is much smaller in magnitude than the response to 1 Sv of freshwater forcing with a similar spatial characterization with an amplification of the SH precipitation peak and reduction of the NH peak. The ensemble mean ΔP response to 0.1 Sv is not significantly different than 10% of the ensemble mean ΔP in response to 1 Sv of hosing in the CCSM4 (12 simulations) with an ensemble mean (southward) optimal shift (S) of $-0.12 \pm 0.03^\circ$ in response to 0.1 Sv as compared to $-1.5 \pm 0.2^\circ$ in response to 1 Sv. This result indicates that the number of simulations considered can not reject the null hypothesis that the ΔP response

to hosing is linear with respect to the magnitude of the prescribed freshwater
flux.

In the 1 Sv freshwater hosing simulations, there is also a modest but robust reduction of the ensemble mean precipitation ($I = 0.97 \pm 0.01$) that explains a small fraction of ΔP (1.2%), whereas the ensemble mean contraction is not robustly different from zero. Despite the ambiguity of the direction of contraction across ensemble members (Fig. 3B), the contracting mode describes a non-negligible fraction of ΔP (4.0%) across the ensemble members and up to 22.5% in individual ensemble members (Fig. 4A). We note that, even though the shifting mode explains a larger fraction ΔP than the other two modes, overall the ΔP is poorly characterized (26.2%) by the geometry of the optimal modes considered here, as is evident in the mismatch between the optimal fit and P_{FORCED} (c.f. the solid and dashed orange lines in Fig. 2B). The reason for this mismatch is as follows: the three modes chosen in this study do not allow for the relative magnitudes of the precipitation peaks in the NH and SH to change but, rather, only allow the peaks to simultaneously increase/decrease (via the intensity mode), translate together (via the shift mode) or move closer/farther apart (via the contraction mode). If we were, instead, to employ a precipitation mode that optimized the intensity of the precipitation differentially in the NH and SH, the optimized fit would better match P_{FORCED} under freshwater forcing. We return to this point in Section 4.3.

4.1.3 Mid-Holocene

The ensemble average ΔP in response to mid-Holocene forcing is small in magnitude compared that of other forcings considered here (green line in Fig. 2B). Furthermore, there is no robust direction of the optimal shift, intensification and, contraction across the ensemble of experiments. For individual ensemble members, the magnitude of simulated ΔP in response to mid-Holocene forcing is significantly smaller (approximately 20% of) that in the other forcings considered here with an ensemble average tropical domain RMS of ΔP value of 0.08 mm/day as compared to 0.37, 0.40 and 0.34 mm/day in the LGM, 4XCO₂ and hosing simulations, respectively. Additionally, Fig. 4A shows that the total fraction of ΔP explained by all three optimal modes is significantly smaller for the mid-Holocene simulations (ensemble average of 25.2%) than for the LGM and 4xCO₂ simulations (ensemble averages of 61.8% and 61.3%, respectively). The contraction mode describes the largest fraction of ΔP across the mid-Holocene ensemble (12.4%) followed by the shifting mode (11.3%) whereas the intensification mode is relatively inefficient at characterizing ΔP (2.4%). We note that there are robust seasonally and longitudinally inhomogeneous changes in simulated tropical precipitation under mid-Holocene forcing (Braconnot *et al.*, 2007b; Liu *et al.*, 2018) but the imprint of these changes on the zonal and annual mean ΔP is small and poorly characterized by the modes of tropical precipitation considered here.

4.1.4 Volcanic forcing

The tropical precipitation response to volcanic forcing is separated into ensemble averages of NH volcanoes (red line - Fig. 2C), SH volcanoes (blue line) and, tropical volcanoes (green line). The ordinate scale for each ensemble average (NH, SH, and tropical) are offset for visual purposes. The ensemble mean P_{FORCED} in response to NH volcanic events features a reduction in the amplitude of the NH precipitation peak and an enhancement of the SH precipitation peak relative to P_{PI} . This precipitation response is qualitatively similar to that seen under North Atlantic freshwater hosing but with reduced magnitude. The response to SH volcanoes is nearly a mirror image of the response to NH volcanoes but with an additional tropical domain average precipitation reduction such that a reduction in the amplitude of the SH precipitation peak is the primary feature of ΔP . ΔP under extratropical volcanic forcing is robustly characterized by an ensemble mean shift away from the source of volcanic forcing: the optimal shift in response to NH volcanic forcing is southward in all 15 ensemble members (ensemble mean $-0.4^\circ \pm 0.2^\circ$) and the optimal shift in response to SH volcanic forcing is northward in all nine ensemble members (ensemble mean $0.4^\circ \pm 0.2^\circ$). In addition to the shift under extratropical volcanic forcing, there is also a robust de-intensification of precipitation under SH volcanic forcing and a contraction of precipitation under NH forcing. These two robust ensemble mean responses (de-intensification under SH forcing, contraction under NH forcing) are most pronounced in the three GISS Model E members that have a larger magnitude volcanic forcing

than the other simulations (see *Atwood et al.*, 2020, for a discussion) whereas the shifting response to extratropical volcanoes is robust across models (Fig. 4A).

The fraction of ΔP explained by the shifting mode across the ensemble members is 17.7% for the NH volcanic events and 12.3% for SH volcanic events. This result suggests that, despite the robust ensemble mean response projecting onto the optimal shifting mode, *the precipitation response to extratropical volcanic forcing is poorly characterized by a meridional translation of the climatological precipitation*. As was discussed for ΔP under freshwater hosing, the shifting mode cannot capture the opposing changes in the magnitude of the NH and SH precipitation peaks and, thus, fails to capture a large fraction of ΔP . The intensifying and contracting modes explain 18.7% (4.0%) and 9.8% (13.2 %) of ΔP across the SH (NH) volcanic forcing indicating that despite the less robust ensemble mean changes in these modes, contractions and intensifications characterize a significant portion of ΔP in individual ensemble members (Fig. 4A).

The ensemble mean P_{FORCED} in response to tropical volcanoes features a nearly latitudinally homogeneous reduction in precipitation throughout the tropical domain with no discernible change in the shape of the distribution (green line in Fig. 2C). The optimal characterization of the ensemble mean ΔP under tropical volcanic forcing is a robust reduction in intensity (6% reduction in the intensity scalar, I) with no statistically significant contraction or shift. Across the ensemble members, the intensification mode explains 37.5%

of ΔP whereas the contraction explains 18.9% despite the differing direction of contraction/expansion across the the ensemble members (contracting in 5 of 9 members and expanding in 4 of 9 members). The shifting mode explains very little (2.3%) of ΔP in response to tropical forcing which suggests that, in the particular case of tropical volcanic forcing, changes in energy input to the tropical atmosphere (*Schneider et al.*, 2014) has minimal impact on the ITCZ location perhaps because the annual mean (mean state) ITCZ is near the equator and, thus, there is minimal area between the energy flux equator and geographic equator.

4.2 The emergence of a a joint contracting and intensifying mode of tropical precipitation

Across all of the different forcing experiments considered, the optimal contraction and expansion scalars are remarkably anti-correlated ($R=-0.91$, Fig. 3B). Additionally, these two modes respond robustly and nearly unanimously across different climate models under the same climate forcing, with a contraction and intensification of P_{FORCED} in response to global warming under $4XCO_2$ and an expansion and de-intensification of P_{FORCED} in response to global cooling during the LGM – as indicated by the LGM and $4XCO_2$ simulations clustering in quadrant II and IV of the contraction/intensification scalar plane respectively (Fig 3B). These results suggest that tropical precipitation contractions and intensifications are tightly coupled and the direction of these forced changes are consistent across models, though the magnitude (of the

combined contraction/intensification) is model dependent. Moreover, the scaling between the contraction (C) and intensification (I) scalars seems relatively consistent across models and climate forcings with a least squares regression slope of $C=-0.8I$ (the dashed black line in Fig. 3B). Given that both C and I project onto tropical domain average precipitation (which increases with both increasing I and decreasing C), it is possible that the correlation between C and I seen in Fig. 3B results from the mutual relationship to domain average precipitation. We test this hypothesis by contracting P_{PI} (via random values of C in equation 5) and finding the associated I required to conserve tropical domain average precipitation which results in the (linear best fit) relationship of $C=-1.95 I$. This tropical precipitation conserving scaling between C and I differs from that seen across the climate forcing simulations and, thus, suggests that the tight correlation between C and I seen has a physical basis and is not a statistical artifact.

Given the tight correlation and consistent scaling between C and I seen across the ensemble of climate forcing simulations, we now define a combined mode of contraction and intensification constrained by the linear best fit relationship of $C=-0.8I$. In other words, we require that C and I used to define the optimal fit to P_{FORCED} in Eq. 5 follow $C=-0.8 I$ such that specifying either C or I constrains the other, thus, reducing the number of free parameters used to optimize the fit to P_{FORCED} . We refer to this combined mode of contraction/intensification or expansion/deintensification as the CI mode. We repeat our optimal characterization ΔP using two modes -- the shift

and *CI* mode — as opposed to the three original independent modes (shift, contraction and, intensification). Since we are fitting for two free parameters instead of three free parameters using the same geometric basis, a smaller fraction of ΔP will be described by the two mode methodology as compared to the three mode methodology. However, the advantage of this approach is it simplifies the characterization of tropical precipitation changes and reduces the likelihood of statistical artifacts.

The fraction of ΔP explained by the two mode characterization across all forcing scenarios is shown in Fig. 4B alongside that of the three mode characterization in Fig. 4A. We note, that since the shift mode is optimized to best fit P_{FORCED} before optimizing the other modes in the calculation, $FRAC_{SHIFT}$ is unchanged between the three mode and two mode characterizations. In the LGM and 4XCO₂ simulations, the *CI* mode explains a very large fraction (48.3% and 53.1%, respectively) of ΔP , respectively. There is a modest reduction (8.7% in the LGM simulations and on 5.4% in the 4XCO₂ simulations) in ΔP explained due to switching from a three parameter characterization to the two parameter characterization (c.f. panel 4A and 4B) suggesting that the *CI* mode efficiently represents ΔP across these simulations (and across the different models) with little loss of explanatory power. Additionally, the precipitation response to tropical volcanoes, and to some extent SH volcanoes, is well characterized by the *CI* mode (which explains 36.2% and 27.5% of ΔP respectively). In contrast, in response to freshwater hosing and NH volcanoes, the *CI* mode explains very little of ΔP , as is expected given that ΔP is

dominated by the shift mode. Under mid-Holocene forcing, combining the contraction and intensification modes into a single CI modes results in reduction of ΔP explained by 8.1%, from 26.1% in the three mode characterization to 18% for the two mode characterization. Similarly, under volcanic forcing the fraction of ΔP explained by the two mode optimization (32.2%) is reduced by 11.1% relative to that of the three mode optimization (43.3%). These significant fractional loss of explanatory power by the two mode optimization results from the optimal contraction and intensification scalars in individual ensemble members diverging from the ensemble mean best fit line (Fig. 3B).

Overall, the CI mode is an efficient representation of the dominant mode of ΔP in the simulations characterized by global warming and cooling (e.g. LGM and $4XCO_2$) and only modestly reduces the explanatory power of the modal characterization of ΔP under forcings that are hemispherically asymmetric and poorly characterized by the precipitation modes considered here (i.e. mid-Holocene, freshwater hosing, volcanoes).

4.3 Magnitude and explanatory power of the shifting mode

The optimal shift (S) defined in this study provides a metric to quantify an ITCZ shift in terms of its most literal definition, that is, a meridional translation of the climatological precipitation pattern. This metric can be compared to other commonly used metrics of forced “ITCZ shifts” such as changes in the precipitation centroid (P_{CENT} , *Frierson and Hwang, 2012*) or changes in the precipitation weighted latitude (“moments”) of the tropical precipitation

distribution (*Adam et al.*, 2016). We present the relationship between S and the P_{CENT} change in Fig. 3A due to the prominence of the latter metric in the paleoclimate (*McGee et al.*, 2014) and dynamics (*Donohoe et al.*, 2013) literature. S and the change in P_{CENT} are very well correlated ($R=0.93$) across the ensemble of different forcings but the magnitude of S is approximately half of the P_{CENT} change (the slope of the linear best fit in Fig. 3A is 0.52) indicating that the P_{CENT} metric of “ITCZ shifts” is more sensitive to the patterns of ΔP realized in model simulations than the optimal shift metric S . Changes in S are generally small in magnitude (ordinate of Fig. 3B): S is less than 1° in all $4XCO_2$, mid-Holocene and volcanic forcing simulations and only exceeds 1° in magnitude for two of the 13 LGM simulations. The largest magnitude S values are found in response to freshwater hosing with magnitudes exceeding 1° in 15 of 18 simulations. S is less than 2° in magnitude across all simulations.

Why are ITCZ shifts calculated by S approximately half the magnitude of those calculated by P_{CENT} changes? The reason has to do with the fact that the forced precipitation changes are poorly described by meridional translations of the climatological precipitation. To see this point, we first analyze the quantitative relationship between P_{CENT} – defined as the latitude that demarcates regions of equal area weighted precipitation between $20^\circ S$ and $20^\circ N$ – and the optimal shift S for the case of a ΔP that is perfectly described by a meridional translation of the climatological precipitation. We do so by meridionally translating P_{PI} via Eq. 1 north and south in increments of 0.1°

up to 5°N/S and calculate the resulting P_{CENT} change for each S . We find that the P_{CENT} change is approximately one half of S for all cases (regression coefficient = 0.51). This suggest that the optimal shift (S) is a more sensitive metric for meridional translations of the ITCZ than the P_{CENT} change metric which can be understood as follows. P_{CENT} (the latitude demarcating equal regions of spatially weighted precipitation) can only change as much as S (the meridional translation of P_{PI}) if the precipitation is zero at the edge of the tropical domain because, otherwise, the precipitation that enters and leaves the tropical domain ($20\text{S}-20\text{N}$) as the precipitation pattern translates will cause P_{CENT} to move less than S . Because precipitation is non-zero on the edge of the tropical domain, one would expect S values to exceed P_{CENT} changes *if ΔP was well characterized by a meridional translation of P_{PI}* which is the opposite of what is seen in our results (Fig. 3A). Instead, our results suggest that S values are less than changes in P_{CENT} because the ΔP changes are so poorly described by meridional translations of the climatological precipitation. Instead, the P_{CENT} changes primarily reflect opposing changes in the amplitude of the NH and SH tropical precipitation peaks (as seen in the ensemble mean ΔP under freshwater hosing; Fig. 2B). Such opposing changes in the NH and SH precipitation peaks strongly impact P_{CENT} but only weakly project on to S .

5 Summary and discussion

The shifting mode (S) of (annual and zonal mean) tropical precipitation changes (ΔP) – defined here as the optimal meridional translation of the unperturbed precipitation – is found to be relatively useless for characterizing simulated zonal mean precipitation changes under a myriad of anthropogenic and paleoclimatic forcings. In response to global scale warming and cooling associated with LGM boundary conditions and CO₂ quadrupling, the fraction of ΔP described by S ($\text{FRAC}_{\text{SHIFT}}$) is 4% and 2% respectively. Even in the cases of strong hemispherically-asymmetric forcings such as freshwater hosing and NH volcanic forcing where $\text{FRAC}_{\text{SHIFT}}$ (19% and 17% respectively) exceeds FRAC_{INT} and FRAC_{CON} , ΔP is poorly characterized by the shift mode. In these cases, ΔP is also poorly characterized by the optimal fit of all tropical precipitation modes considered here with only 25% and 31% of the total ΔP explained as compared to 59% for all other simulations. Hemispherically asymmetric tropical precipitation responses are best characterized by changes in the relative amplitude of the tropical precipitation peaks in the NH and SH. These precipitation changes are distinct from an ITCZ shift, in its most literal sense.

It is possible that our interpretation of the term “ITCZ shift” as a meridional translation of the climatological precipitation pattern is overly literal and is different from that intended by the paleoclimate and climate dynamics communities. However, our main point here is that a meridional dipole in the pattern of forced precipitation change is not necessarily indicative of (i.e.

statistically well described by) a meridional shift of the climatological precipitation. We hope our work here motivates future studies to clarify the definition of an “ITCZ shift” and quantify how consistent tropical precipitation changes are with the definition used. The various metrics used in literature to quantify “ITCZ shifts” should be renamed to more accurately represent the patterns of forced precipitation changes and the underlying links to the changes in the Hadley circulation that the metrics respond to.

We find that the magnitude of ITCZ shifts quantified by the optimal shift (S) method are generally small – less than 1° latitude in magnitude with the exception of the highly hemispherically asymmetric North Atlantic freshwater forcing (where $|S|$ ranges from 0.4 to 1.7° in response to 1 Sv freshwater hosing). S calculated from the simulated precipitation response to forcing is smaller (by approximately one half) than that inferred using alternative statistical metrics of “ITCZ shifts” (*Adam et al., 2016*) such as P_{CENT} (*Frierson and Hwang, 2012; Donohoe et al., 2013*). Our results provide additional support for the notion that zonal mean ITCZ shifts of more than a couple degrees latitude are not realized in forced model simulations. Small magnitude ITCZ shifts have also been deduced from paleoclimate proxy sea surface temperature (SST) reconstructions during the LGM, mid-Holocene and across Heinrich events using statistical relationships between the inter-hemispheric SST gradient and the zonal mean ITCZ location derived from modern day observed variability and forced simulations (*McGee et al., 2014*). Compilations of paleo hydrological proxy records throughout the tropics meanwhile have

been interpreted as representing larger magnitude ($> 5^\circ$) regional shifts of the ITCZ during the LGM and Little Ice Age (*Arbuszewski et al.*, 2013; *Pahnke et al.*, 2007; *Haug et al.*, 2001; *Sachs et al.*, 2009). Simulated ITCZ shifts in response to paleoclimate forcing are highly zonally inhomogeneous (*Atwood et al.*, 2020). Together, these results suggest that sparse compilations of proxy records likely can only inform on regional (as opposed to zonal mean) shifts of the ITCZ. Our work here further suggests that, even if we had reliable proxy records for reconstructing *zonal mean precipitation*, such changes should not be interpreted solely in terms of a meridional translation of the climatological precipitation the given the small fraction of simulated ΔP explained by the shift mode relative to the intensity and contraction modes. Indeed, model simulations suggest that a decrease (increase) in precipitation proportional to the mean state precipitation throughout the tropics in response to global mean cooling (warming) is a good zeroeth order hypothesis, as trivial as that might sound.

The dynamics literature (*Donohoe et al.*, 2013; *Donohoe and Voigt*, 2015) finds that a zonal mean ITCZ shift (measure by changes in P_{CENT}) of 3° would require a sustained hemispheric asymmetry of atmospheric heating equivalent to a simultaneous doubling of CO_2 in one hemisphere and CO_2 halving in the opposing hemisphere (resulting in 1 PW of atmospheric energy transport across the equator) assuming no radiative feedbacks (which generally damp localized forcing). We cannot comprehend a forcing that could achieve a hemispheric asymmetry in atmospheric heating of the magnitude necessary to shift

the ITCZ by more than a couple of degrees – even more so when the damping impact of ocean circulation anomalies is considered (*Green et al.*, 2019). The preponderance of modeling, theoretical and empirical observational evidence suggests that large magnitude ($> 3^\circ$) zonal mean ITCZ shifts under paleoclimate forcing are dynamically unrealistic and are unlikely to explain the spatial structure of tropical precipitation changes.

We find that contractions and intensifications of the tropical precipitation respond robustly and in unison (with fixed scaling) to external forcing: e.g. tropical precipitation intensifies and meridionally contracts under global warming, and reduces and meridionally expands under global cooling. The tight (empirically derived) scaling between contractions and intensifications allow us to define a joint contracting and intensifying mode (*CI*) mode of tropical precipitation that efficiently describes the majority of the tropical precipitation changes under LGM and $4\times\text{CO}_2$ forcing.

What mechanism accounts for the tight coupling between contraction and intensification seen across the ensemble of forcing experiments (Fig. 3B)? The meridional contraction of tropical precipitation in response to global warming under anthropogenic forcing has been noted in several studies (*Lau and Kim*, 2015; *Su et al.*, 2017). The proposed mechanisms of precipitation contraction under warming include: (i) changes in the efficiency of transient-eddy energy and moisture fluxes (*Byrne and Schneider*, 2016a; *Nolan et al.*, 2010; *Peters et al.*, 2008), (ii) cloud radiative effects (*Su et al.*, 2017), and (iii) increased gross moist stability (GMS) of the tropics under warming that impacts the

meridional width of convective regions (*Chou et al.*, 2009; *Byrne and Schneider*, 2016b). *Donohoe et al.* (2019) found that a contraction of annual mean precipitation resulted from a reduced seasonal migration of the ITCZ off the equator under global warming and argued that the latter resulted from enhanced GMS of the tropical atmosphere. All of these mechanisms would be expected to lead to an expansion of tropical precipitation under global cooling in addition to the contraction seen under global warming as found in the present study. However, it is unclear whether the degree of simulated contraction is controlled by the magnitude of tropical surface warming, the GMS change, the degree of precipitation intensification (which would be consistent with our results in Fig. 3B), and/or by changes in seasonality. In future work, we hope to apply the optimal characterization of ΔP to seasonal precipitation changes to disentangle these proposed mechanisms by asking whether annual mean contractions result from contractions during all seasons or, alternatively, reflect seasonal precipitation shifts during the solstitial seasons that rectify to an annual contraction. Given the strong seasonality of zonal mean tropical precipitation, we also plan to examine whether the large fraction of ΔP explained by the *CI* extends to seasonal precipitation changes or, rather, the annual mean changes reflect more subtle changes in the seasonality of precipitation.

How can our findings here be used to interpret large-scale patterns of past tropical precipitation changes based on compilations of sparsely located paleoproxy hydroclimate records? One possible path forward is to shift, contract and intensify the observed climatological precipitation pattern (*Xie and Arkin*,

1996) to optimally fit (i.e. minimize the RMS mismatch) the discrete proxy data. If we can trust the scaling between the contraction and intensification scalars found across the ensemble of climate model simulations (Fig. 3B), then we can reduce the number of free parameters in the optimal fit to two (the shift and *CI* mode), thus, improving statistical significance of the fit to proxy data. This procedure would allow the optimal shift and *CI* mode change to be quantified in past climates.

Our work here has focused exclusively on zonal and annual mean tropical precipitation changes whereas the simulated tropical precipitation response has been shown to feature strong zonal inhomogeneities (*Braconnot et al.*, 2007a, 2012) with robust zonally localized responses to forcing (*Atwood et al.*, 2020). We plan to extend our optimal shifting, intensifying and contracting framework to characterize zonal asymmetries in the precipitation response to forcing in order to better understand and relate mechanisms of regional precipitation changes to compilations of proxy records.

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Fig. 1 Illustration of optimal fitting to shifted/contracted/intensified modes of tropical precipitation and the procedure used to define the percent of the precipitation change explained by each mode. The black line shows the climatological (zonal and annual mean) precipitation in the control (pre-industrial) simulations and the green line shows the precipitation after external forcing. First the precipitation is optimally shifted (P_{SHIFT} , blue line). Next the precipitation is optimally intensified and shifted together ($P_{SHIFT,INT}$, red line). Finally, the precipitation is optimally contracted, intensified and shifted together ($P_{SHIFT,INT,CON}$, orange line). The percentage of ΔP explained by each mode is defined as the additional reduction in (normalized) RMS ΔP by successively introducing the mode and all optimal parameters are defined from optimizing the three modes simultaneously. Panel (A) shows an example of a $4XCO_2$ precipitation response that is well described by a intensification and contraction. Panel (B) shows the precipitation response to freshwater hosing that is best described by a southward shift.

Fig. 2 Ensemble mean, zonal mean and annual mean precipitation for pre-industrial simulations (solid black) and forced simulations (solid colors) with the optimal fit to the forced precipitation shown in dashed colors for the: (A) Last Glacial Maximum (blue) and 4XCO₂ (red) simulations; (B) Mid-Holocene (green) and 1 Sv North Atlantic Freshwater Hosing (orange) and (C) Volcanic forcing simulations subdivided into eruptions originating in the Northern Hemisphere (red), Southern Hemisphere (blue) and tropics (green). Results from each volcanic forcing region have been offset on the ordinate by 2 mm per day (with the PI distribution repeated) for visual purposes as indicate by the colored ordinate axes. Ensemble mean optimal parameters and their inter-model spread (defined as 2 standard deviations of the mean) are shown in the inset tables.

Fig. 3 Optimal shift, contraction and intensification parameters for the zonal and annual mean precipitation changes in response to LGM (blue dots), 4XCO₂ (red dots), Mid-Holocene (green dots), freshwater hosing (orange dots for 1 Sv and crosses for 0.1 Sv) and volcanic forcing which are further subdivided into volcanoes in the Northern Hemisphere (NH – red crosses), Southern Hemisphere (SH – blue crosses) and, Tropics (TROP – green crosses). (A) Optimal shift (ordinate) versus the change in precipitation centroid (abscissa) that is commonly used to quantify ITCZ shifts. The 1:1 line is shown by the solid black line and the linear best fit is shown by the dashed black line. (B) Optimal contraction scalar (ordinate) versus optimal intensification scalar (abscissa). Dots (and smaller crosses for volcanoes) represent individual models and the open squares (larger plus signs for volcanoes) show the ensemble mean of simulations for each forcing experiment. The ensemble mean square for the hosing simulations is the mean of the 1 Sv freshwater forcing simulations only.

Fig. 4 The percentage of ΔP explained by each mode of tropical precipitation changes in the ensemble of forced simulations. Results are grouped by forcing type with individual models shown by thin bars and the ensemble average shown in the thick bar to the right. (A) Three mode optimization with percentages defined starting from the shifting mode only (blue) then adding the intensification mode (red) and the adding the contraction (orange). (B) Two mode optimization with percentages defined starting from the shifting mode (blue) then adding the combined contraction-intensification mode (CI—reddish orange) with the ratio of the contraction to intensification defined from the linear best fit across all simulations shown in Fig. 1B.