Spatial Patterns of Precipitation Change in CMIP5: Why the Rich Do Not Get Richer in the Tropics

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ABSTRACT

Changes in the patterns of tropical precipitation ($P$) and circulation are analyzed in Coupled Model Intercomparison Project phase 5 (CMIP5) GCMs under the representative concentration pathway 8.5 (RCP8.5) scenario. A robust weakening of the tropical circulation is seen across models, associated with a divergence feedback that acts to reduce convection most in areas of largest climatological ascent. This is in contrast to the convergence feedback seen in interannual variability of tropical precipitation patterns. The residual pattern of convective mass-flux change is associated with shifts in convergence zones due to mechanisms such as SST gradient change, and this is often locally larger than the weakening due to the divergence feedback.

A simple framework is constructed to separate precipitation change into components based on different mechanisms and to relate it directly to circulation change. While the tropical mean increase in precipitation is due to the residual between the positive thermodynamic change due to increased specific humidity and the decreased convective mass flux due to the weakening of the circulation, the spatial patterns of these two components largely cancel each other out. The rich-get-richer mechanism of greatest precipitation increases in ascent regions is almost negated by this cancellation, explaining why the spatial correlation between climatological $P$ and the climate change anomaly $\Delta P$ is only 0.2 over the tropics for the CMIP5 multimodel mean. This leaves the spatial pattern of precipitation change to be dominated by the component associated with shifts in convergence zones, both in the multimodel mean and intermodel uncertainty, with the component due to relative humidity change also becoming important over land.

1. Introduction

Tropical rainfall is projected to undergo significant changes in both magnitude and spatial pattern in its response to greenhouse gas forcing (Meehl et al. 2007). This has the potential to produce substantial climate impacts, particularly in countries already vulnerable to interannual and decadal variability in rainfall. The magnitude of global mean precipitation increase per degree of warming is relatively well constrained in GCM estimates by the approximate balance between changes in net tropospheric radiative cooling and latent heating (Lambert and Webb 2008). However, the level of agreement between GCMs on regional precipitation changes is low throughout large areas of the tropics, with even the sign of change uncertain for many regions (Meehl et al. 2007; Rowell 2012). Therefore, the mechanisms that lie behind the spatial patterns of the tropical rainfall response to climate change are of great interest.

The aim of this study is to investigate these mechanisms and to better understand how they combine to produce both the mean pattern of precipitation change in the Coupled Model Intercomparison Project phase 5...
(CMIP5) and the variation in this pattern across models. The precipitation response to warming can be thought of as a combination of thermodynamic and dynamic changes, and several previous studies have used this framework to examine precipitation change in CMIP3 (Vecchi and Soden 2007; Seager et al. 2010; Muller and O’Gorman 2011; Chou and Neelin 2004; Chou et al. 2009). The rich-get-richer hypothesis (Held and Soden 2006; Chou et al. 2009), associated with thermodynamic increases in moisture transport, would lead to wet regions becoming wetter and dry regions becoming drier in the absence of compensating circulation changes. However, this proves a poor description of the spatial pattern of precipitation \((P)\) change within the tropics \((30^\circ N\text{–}30^\circ S)\), with the spatial correlation between the climate change anomaly \(\Delta P\) and climatological \(P\) being 0.2 for the multimodel mean and a maximum of 0.3 across the CMIP5 models (for the difference between the periods 1971–2000 and 2071–2100 under the RCP8.5 scenario). Therefore, it appears likely that dynamical precipitation changes are acting in some way to counteract the pattern of thermodynamic rainfall change.

The first part of this study examines changes in the tropical circulation in CMIP5. The insight gained into circulation change is then used to construct a novel decomposition of precipitation change, allowing the various dynamic and thermodynamic mechanisms to be examined and their relative contributions quantified.

Previous analyses of tropical circulation and rainfall changes under warming in GCMs have yielded a number of interesting findings. Held and Soden (2006) used the CMIP3 archive to look for robust responses of the water cycle to global warming. One of their key results was a general weakening of the tropical circulation, which they explained by making the approximation that

\[
P = Mq,
\]

where \(P\) is precipitation, \(M\) is mass flux from the boundary layer to the free troposphere, and \(q\) is a measure of boundary layer specific humidity. As global mean \(q\) increases in GCMs at approximately the Clausius–Clapeyron rate of 7% \(K^{-1}\), whereas a limited increase in tropospheric radiative cooling leads global mean \(P\) to increase at only 1–3% \(K^{-1}\) (Mitchell et al. 1987; Stephens and Ellis 2008), \(M\) is constrained to decrease with warming.

A general weakening of convective mass flux thus implies a weakening of the tropical circulation (at least in the absence of a wholesale reorganization of the circulation that is not seen in GCMs). An observed weakening of the Walker circulation both over the twentieth century (Vecchi et al. 2006) and over the last six decades (Tokinaga et al. 2012), although recently questioned by Meng et al. (2011), provides evidence that the signal of tropical circulation change may already be emerging from natural variability. Other studies (Bosilovich et al. 2005; Li et al. 2011) have examined the related concept of the atmospheric moisture recycling rate in GCMs and observations and conclude that this is increasing, as expected from a weakening circulation.

Although Held and Soden (2006) provide a simple and seemingly robust constraint on tropical circulation change, this by itself does not lead to great insight into the physical mechanisms behind the weakening. Knutson and Manabe (1995) proposed such a mechanism, noting that in subsidence regions radiative cooling does not increase as fast as dry static stability under greenhouse gas forcing. This implies a decrease in the rate of radiatively driven descent and a weakening of the descending branch of the tropical circulation.

Recently, Ma et al. (2012) have shed light on the reduction of pressure gradients between ascent and descent regions under warming that is associated with a weakening of the divergent winds. They used the Geophysical Fluid Dynamics Laboratory (GFDL) Climate Model version 2.1 (CM2.1) to show that total tropospheric column warming is greater in climatological descent regions than in ascent regions under both greenhouse gas forcing and uniform sea surface temperature (SST) increase experiments. This is in contrast to events such as El Niño, where localized latent heating leads to a local column temperature increase, producing horizontal pressure gradient changes and a convergence feedback at low levels (An and Wang 2001). Ma et al. (2012) explain this difference between greenhouse gas forcing and natural variability by the different ratios of global mean SST warming to spatial variance of SST warming in the two cases. Under greenhouse gas forcing, the global mean SST warming is larger than the spatial variance, whereas the converse is true for natural variability. This leads to the mean advection of stratification change (MASC) becoming important under global warming, producing greater column warming of the descent regions than the ascent regions, and reducing pressure gradients between them.

CMIP3 models show a greater weakening of the Walker circulation than the Hadley cells (Vecchi and Soden 2007), which is explained by Gastineau et al. (2009) and Ma et al. (2012) as the effect of the pattern change in latent heating due to SST pattern changes and variation across models of these SST changes. The importance of SST pattern changes was also highlighted by Xie et al. (2010), who suggested that they are the dominant influence on the pattern of future precipitation change in the tropics.
In contrast to these results showing a weakening of the tropical circulation, Hoyos and Webster (2012) found that the response of a simple two-level nonlinear zonally symmetric model to SST warming was to increase the strength of the tropical circulation, and they hypothesized that the tropical circulation should strengthen under warming to compensate for the increased latent heating in the ascent regions. However, it seems probable that this two-level model is unable to properly represent the mechanisms simulated in GCMs (e.g., by failing to vertically resolve the MASC mechanism) and therefore exhibits a response similar to the convergence feedback of natural variability rather than the weakening of the circulation seen in GCMs.

It is not obvious how this weakening of the tropical circulation will manifest itself spatially and how it would interact with other mechanisms of precipitation change such as the rich-get-richer mechanism. Chou et al. (2009) proposed that the weakening is the result of a combination of the “upped ante” mechanism reducing precipitation on convective margins, combined with a weakening of convection in some ascent regions because of a raising of the convective outflow height at upper levels (Chou and Chen 2010). However, the reason why these mechanisms would apply selectively in some regions but not in others is unclear. This hypothesis can be seen as a bottom-up view of the weakening of the tropical circulation, whereby various separate mechanisms acting in different regions combine to satisfy the constraint of Eq. (1), as opposed to a top-down view consistent with Ma et al. (2012), where a single dynamical mechanism acts across the tropics to slow down the circulation.

Seager et al. (2010) provided a thorough global analysis of the contribution to changes in $P - E$ (precipitation minus evaporation) by thermodynamic and dynamic components, but they did not separate out changes in precipitation from those in evaporation or dynamical changes related to the weakening of the tropical circulation from those attributed to other factors. Muller and O’Gorman (2011) used an energy budget approach to analyze changes in regional precipitation and decomposed the change into thermodynamic and dynamic components, but again, they did not specifically isolate the component of change due to the weakening of the tropical circulation.

In this study, data from the CMIP5 archive are used to examine changes in the tropical circulation under greenhouse gas forcing and how these are linked to changes in the spatial pattern of convective mass flux. A simple framework is then proposed to relate these circulation changes to the spatial patterns of precipitation change in the tropics and to illuminate the mechanisms behind them. In particular, the direct contribution of the weakening of the tropical circulation to the pattern of tropical precipitation change is isolated. The area of interest is restricted to the tropics, where the majority of rainfall is convective, allowing circulation changes to be examined through changes in convective mass flux.

Section 2 describes the data used in this study, followed by analysis of changes in the mean tropical circulation and associated patterns of convective mass flux in sections 3 and 4. Section 5 describes the framework used to relate spatial changes in mass flux to precipitation changes, and the results of this analysis are shown in section 6. Finally, these results are summarized and discussed in section 7.

2. Data

Monthly mean data from 14 CMIP5 models for the historical and representative concentration pathway 8.5 (RCP8.5) experiments were downloaded from the Earth System Grid and averaged to annual means. At the time of writing, not all variables needed for this study were available from all CMIP5 models, so only this subset was used. To compare the models at the same spatial scale, all data were regridded to a common resolution of 2.5°. Convective mass-flux data from two of the models were found to be in error, with the correct variable not having been saved by the modeling group, so data from the remaining 12 models (listed in Table 1) were used in this study. For models where more than one ensemble member was created for the RCP8.5 scenario, only the first member is used. In the appendix, diagnostics from the HadGEM2-ES RCP8.5 run that are not available from the CMIP5 archive were used to examine the tropical moisture cycle.

3. Tropical mean convective mass-flux change in CMIP5

We begin our analysis of tropical circulation changes in CMIP5 by examining changes to the tropical mean circulation. As convection is parameterized in GCMs, and convective rainfall dominates over large-scale rainfall at low latitudes, within the tropics Eq. (1) is primarily a constraint on convective mass flux ($M_c$). Vecchi and Soden (2007) showed that global mean $M_c$ at 500 hPa decreases in GFDL CM2.1 under the Special Report on Emissions Scenarios A1B scenario. As $M_c$ was not available in the CMIP3 archive, they derived an estimated measure of mass flux $M'$ from Eq. (1) for the CMIP3 models: $\Delta M'/M' = \Delta P/P - 0.07 \Delta T$, where $T$ is air temperature 2 m above ground (2-m temperature) and $\Delta$ indicates the difference from the end of the twentieth-century climatology for each variable. Global mean $M'$
weakens under the A1B scenario for all CMIP3 models and so was taken as a measure of robustness of the weakening of the tropical circulation in CMIP3.

In CMIP5 $M_c$ is available for all models and is examined here (the variable used is actually updraft minus downdraft convective mass flux). Although $M_c$ is a parameterized quantity in GCMs, it is strongly correlated with the resolved vertical velocity ($\omega$) (Vecchi and Soden 2007) and so can be used as a measure of the strength of the large-scale resolved tropical circulation. The advantage of using $M_c$ rather than $\omega$ is that it can be directly related to changes in GCM convective rainfall.

The tropical mean change in $M_c$ at 500 hPa for CMIP5 models under the RCP8.5 scenario is shown in Fig. 1a. For the purposes of this study, the tropics are defined as 30°N–30°S (the sensitivity of this study to the choice of these bounds is examined in section 4). Also shown in Fig. 1b is the change in the tropical mean of a proxy measure of mass flux derived from Eq. (1):

\[ M^* = \frac{P}{q}, \]

where $q$ is 2-m specific humidity and $P/q$ is calculated at each grid point before averaging to the tropical mean. $M^*$ is very similar to the $M^*$ of Vecchi and Soden (2007) but does not assume constant relative humidity (RH) and will prove more useful than $M^*$ in the examination of different components of precipitation change contained in section 5.

It can be seen that although $M^*$ weakens in all CMIP5 models, $M_c$ at 500 hPa displays a more diverse range of responses, with several models showing no weakening trend. This does not in fact indicate the lack of a robust weakening of the tropical circulation in CMIP5, but is instead due to the rise in height of convective outflow with warming. Figure 2 shows the change in tropical mean vertical profiles of $M_c$ for each of the CMIP5 models, showing a robust increase in the depth of convection as well as a clear reduction in total mass flux. This serves to stretch the profiles of $M_c$, causing the change in $M_c$ on a single level such as 500 hPa to reflect both the weakening of the circulation and the change in vertical profile.

Following Chadwick et al. (2013), the vertically integrated $M_c$ from the surface to 30 hPa (referred to

\[ M_{int} = \int_{30\ hPa}^{500\ hPa} \frac{P}{q}, \]
hereafter as $M_{int}$) is used as a more appropriate indicator of the change in strength of convection for high forcing scenarios where the change in depth of convection is significant. Sensitivity tests showed that integrating to any level above 200 hPa made very little difference to the value of $M_{int}$. Similarly, integrating upward from 850 hPa instead of the surface so as not to include the shallowest convective updrafts made little qualitative difference to $M_{int}$, so the integral from the surface was used.

Figure 1c shows the tropical mean change in $M_{int}$ under the RCP8.5 scenario, and the CMIP5 models exhibit a robust weakening of the tropical circulation. This weakening occurs for all models examined here, even though there is significant diversity in the shape of the GCMs’s climatological vertical mass-flux profiles (see Fig. 2) and spatial distributions of mass flux. As noted by Chadwick et al. (2013) for HadGEM2-ES, in CMIP5 the percentage change in $M^*$ is larger than that of $M_{int}$ (by on average 50%) for all but two models where the changes are approximately equal. The relationship between $M_{int}$ and $M^*$ is examined in detail in the appendix, and we now discuss how changes in $M_{int}$ are spatially distributed in the tropics.

4. Spatial patterns of mass-flux change

Convection is strongly related to low-level moisture convergence, and any changes in convergence resulting from the weakening of the tropical circulation will be reflected in the spatial pattern of convective mass-flux change. The MASC mechanism of Ma et al. (2012) predicts changes in convergence under warming related to the circulation slowdown, and these can be examined using the CMIP5 models. In this section, we consider whether the projected patterns of change in CMIP5 mass flux are consistent with the predictions of MASC and how this might combine with other changes in the patterns of convection under warming.

Ma et al. (2012) forced a linear baroclinic model with a term that represents MASC, derived from GFDL...
CM2.1 under the A1B scenario, and which is inversely proportional to the climatological vertical wind field. The resultant upper-level velocity potential anomaly is shown in Fig. 6a of Ma et al. (2012) and is approximately, though not exactly, inversely proportional to the climatological velocity potential $\chi$, so $\Delta \chi \approx -a \chi$, where $a$ is constant. This leads to the change in divergence $\Delta \delta$ following the same relationship:

$$\Delta \delta \approx -a \delta,$$  \hspace{1cm} (3)

where $\delta$ is the climatological divergence. Although the 850 hPa velocity potential is not shown in Ma et al. (2012), the low-level velocity potential and divergence fields would be expected to be approximately equal and opposite in sign to the upper-level fields.

This is a divergence feedback on precipitation and latent heating increases under warming (a negative feedback), caused by the weakening of the tropical circulation via the MASC mechanism, with areas of largest climatological low-level convergence experiencing the greatest decreases. It stands in contrast to the convergence feedback that operates in tropical variability and had been thought to play a similar role in climate change–related precipitation pattern change (Chou et al. 2009). The difference between the two can be understood within the MASC framework by the relative importance of the mean tropical stratification change between global warming and tropical variability, and the different gradients of total column temperature change between ascent and descent regions that result in the two cases.

To see whether this divergence feedback can explain any of the spatial pattern of mass-flux change in the CMIP5 models, Fig. 3 shows the normalized change in $M_{\text{int}}$ per degree Kelvin of global mean warming ($\Delta M_{\text{int}}$) plotted against climatological $M_{\text{int}}$ for each grid point for each of the CMIP5 GCMs and the multimodel mean. Gray lines are least square fits to the data. Note that the scale varies between models.
Across all models there is a tendency for greater decreases in convection in regions of larger climatological ascent, as predicted by the MASC mechanism. This is highlighted by the linear least squares fit lines, which indicate some form of inverse relationship between the spatial patterns of $\Delta M_{\text{int}}$ and $M_{\text{int}}$. However, these fits clearly fail to explain many of the features shown in Fig. 3, in particular, the arclike features that contribute the largest positive and negative values of $\Delta M_{\text{int}}$.

Figure 4a shows the multimodel mean change in $M_{\text{int}}$ within the tropics. The largest values are found in the tropical Pacific, with an increase in $M_{\text{int}}$ along the equator and decreases to the north and south in the intertropical convergence zone (ITCZ) and South Pacific convergence zone (SPCZ) regions. There is variability across models as to the exact shape of these features, particularly in their longitudinal distribution within the tropical Pacific, but this large-scale tripole of $M_{\text{int}}$ change is common to all the CMIP5 models used here (not shown). Figure 4d shows the multimodel mean tropical surface temperature change, and a signal of locally enhanced warming along the central and eastern equatorial Pacific can be clearly seen. This is the enhanced equatorial warming identified as a robust feature in CMIP3 models by Liu et al. (2005). Xie et al. (2010) examined the mechanisms behind this feature and hypothesized that enhanced equatorial warming anchors the band of increased rainfall seen across the equatorial Pacific in CMIP3.
It appears likely that the tropical Pacific $M_{\text{int}}$ anomalies of Fig. 4a consist of a shift of convection into the equatorial Pacific from the surrounding regions linked to spatial changes in low-level convergence associated with the local pattern of SST change. To explore these changes in more detail, we focus on two Pacific regions, one of positive $M_{\text{int}}$ anomalies along the equator (region 1: 5°N–5°S, 150°–90°W) and another of negative $M_{\text{int}}$ anomalies south of the equator (region 2: 10°–20°S, 150°–90°W). These regions are shown in Fig. 4a. Figure 5a shows a simplified schematic of these $M_{\text{int}}$ changes represented as a shift of a one-dimensional longitudinal mass-flux anomaly with a sinelike profile from region 2 to region 1. Figure 5b is a schematic of how this idealized shift would appear on a plot of $\Delta M_{\text{int}}$ against $M_{\text{int}}$ such as Fig. 3, and this produces arclike features such as are seen in the GCM data. Although the idealized anomalies are directly latitudinally aligned in Fig. 5a, because of the difference in climatological $M_{\text{int}}$ between region 1 and region 2, the positive and negative $\Delta M_{\text{int}}$ anomalies are offset from each other in the mass-flux space of Fig. 5b.

Figure 6 highlights the $\Delta M_{\text{int}}$ anomalies in regions 1 and 2 for each of the CMIP5 models. Although not conforming exactly to the idealized patterns of Fig. 5b, the behavior of the GCMs in these regions is reasonably well described by this simple model of spatial shifts in convection. As the CMIP5 data have been regridded to 2.5° resolution, this serves to partially discretize the two-dimensional mass-flux anomalies and makes each latitude band of grid points appear to behave somewhat like a separate one-dimensional anomaly. As both region 1 and 2 consist of four such 2.5°-latitude bands, four separate arcs of $\Delta M_{\text{int}}$ can be identified in many of the models in Fig. 6 for each of these regions.

As well as the large shifts of convection in the tropical Pacific, there are likely to be many other shifts in convergence and convection throughout the tropics, associated locally or remotely with SST pattern changes, land–sea temperature gradient changes, land surface changes, and other mechanisms, such as the upslope hypothesis of Chou and Neelin (2004), or other changes in atmospheric dynamics.

Returning to the MASC divergence feedback and the inverse relationship between $\Delta M_{\text{int}}$ and $M_{\text{int}}$ shown by the best fit lines in Fig. 6, we hypothesize that the weakening of the tropical circulation is associated with reductions in $M_{\text{int}}$ that are inversely proportional to the climatological $M_{\text{int}}$. This combines with spatial shifts in convection, which can be locally large compared to the weakening signal and result in significant scatter around the linear inverse relationship, to produce the total pattern of $M_{\text{int}}$ change.

A component of $M_{\text{int}}$ change directly linked to the weakening of the tropical circulation, $\Delta M_{\text{div}}$, can be defined:

$$\Delta M_{\text{div}} = -\alpha M_{\text{int}},$$

where $\alpha$ is constant and strongly related (and possibly equal) to the $a$ of Eq. (3).

Under this framework, the residuals of $\Delta M_{\text{int}}$ from $\Delta M_{\text{div}}$ are the result of spatial changes in convection:

$$\Delta M_{\text{spat}} = \Delta M_{\text{int}} - \Delta M_{\text{div}}.$$  

Here $\Delta M_{\text{spat}}$ is associated with SST pattern changes and any other mechanisms that can result in shifts in low-level convergence, such as the ones described above. Any deviation of the divergence feedback from the linear relationship of Eq. (4) [expected from the conversion of approximately equal ($\approx$) in Eq. (3) to equal ($=$) in Eq. (4)] will also implicitly be included in $\Delta M_{\text{spat}}$. Equation (4) is equivalent to saying that the fractional decrease in convective mass flux due to the weakening of the tropical circulation and associated divergence feedback $\Delta M_{\text{div}}/M_{\text{int}}$ is constant throughout the tropics.

In contrast to this reduction in mass flux in ascent regions, the implication of a divergence feedback in descent regions is that low-level divergence should decrease, potentially leading to a general increase in convection in these regions. The fact that this does not happen in GCMs is likely to be due to the increased dry static stability in descent regions (Knutson and Manabe 1995) inhibiting convection. However, it is possible that the linear relation of Eq. (4) may turn out to need modification in the descent regions.

This hypothesis of the tropical circulation weakening through a constant fractional decrease in mass flux...
across the tropics appears to fit more naturally with the top-down view of a single dynamical mechanism such as MASC driving the weakening of the circulation across the tropics in a unified manner, rather than several different mechanisms in different regions combining together to produce the weakening.

To determine the value of $\alpha$ for each model, one method would be to take the gradient of the least squares fit lines in Fig. 6. However, this would make the assumption that the residuals, $\Delta M_{\text{spat}}$, are symmetrically distributed about zero for each value of $M_{\text{int}}$, but, because many of the largest spatial changes involve shifts from higher to lower areas of climatological $M_{\text{int}}$ in the tropical Pacific, this assumption does not hold. Therefore, the least squares fit lines are likely to be skewed by these large asymmetric $\Delta M_{\text{spat}}$ changes.

An alternative method of determining $\alpha$ is to use tropical mean values of $M_{\text{int}}$ and $\Delta M_{\text{int}}$. Making the approximation that the tropics consist of a closed circulation (this assumption is examined in the appendix), spatial shifts in convection will average to zero at this scale, meaning that $\Delta M_{\text{trop}} = \Delta M_{\text{int}}$ (where the superscript $\text{trop}$ indicates the tropical mean). Therefore from Eq. (4), $\Delta M_{\text{trop}} = -\alpha M_{\text{trop}}$, and the value of $\alpha$ can easily be calculated.

As the Hadley cells expand poleward in GCMs at a rate of around $0.6^\circ$ latitude K$^{-1}$ of warming (Lu et al. 2007; Frierson et al. 2007), calculating $M_{\text{trop}}$ over the same domain for both 1971–2000 and 2071–2100 might be inappropriate. A sensitivity test was performed where $\alpha$ was calculated using a domain of $35^\circ$–$35^\circ$ for both time periods. This widening is larger than that expected from GCMs without a warming of around 9 K. The maximum change to $\alpha$ across the models (compared to that obtained using the original domain of $30^\circ$–$30^\circ$) was around 10%, so even for very large changes in the

**Fig. 6.** As in Fig. 3, but with different regions highlighted in color. Blue crosses indicate grid points in the equatorial Pacific box of region 1 ($5^\circ$–$5^\circ$, $150^\circ$E–$90^\circ$W). Red crosses indicate grid points in region 2 ($10^\circ$–$20^\circ$, $150^\circ$E–$90^\circ$W). Cyan lines are least squares fits to the data. Yellow lines are calculated using the value of tropical mean $\alpha$ determined for each model from Eq. (4). Note that the scale varies between models.
width of the tropics, the effect on $\alpha$ is not large and can reasonably be ignored for the smaller changes likely to be seen in RCP8.5 to 2100. Therefore, the results of mean changes to the tropical circulation shown in Fig. 4 should also be relatively insensitive to the choice of tropical domain.

As can be seen from Fig. 6, in most cases the values of $\alpha$ calculated using the tropical mean are similar to the gradients of the least squares fit lines. Values of $\alpha$ for each model are shown in Table 1. The value $\Delta M_{\text{int}}$ can now be divided into its components $\Delta M_{\text{div}}$ and $\Delta M_{\text{spat}}$ at each grid point, and the results of this decomposition for the multimodel mean are shown in Figs. 4b and 4c.

The pattern of $\Delta M_{\text{spat}}$ over the oceans is largely consistent with the pattern of SST gradient change seen in Fig. 4d. The largest SST increases in each region are collocated with increases in $\Delta M_{\text{spat}}$ and vice versa, which lends weight to the hypothesis that SST gradient changes play a dominant role in determining changes in low-level convergence under warming (Xie et al. 2010).

Although the tropical mean value of $\Delta M_{\text{spat}}$ is zero by construction, the spatial pattern of $\Delta M_{\text{int}}$ is highly influenced by it, with $\Delta M_{\text{div}}$ serving to balance changes in $\Delta M_{\text{spat}}$ in some regions (such as India) and accentuating them in others (such as South America). Thus, the direct effect of the weakening of the mean tropical circulation on the spatial pattern of mass-flux change is in many regions only secondary to the spatial patterns associated with other mechanisms. However, this is not the whole story as, for example, the weakening of the circulation could lead to evaporatively or cloud-change-driven SST changes that would contribute to $\Delta M_{\text{spat}}$. Therefore, although $\Delta M_{\text{div}}$ is independent of $\Delta M_{\text{spat}}$, the converse is not true.

This picture of spatial changes in convergence patterns superimposed on a general weakening of the circulation is not inconsistent with a change in size of the ascent regions compared to the descent regions. However, $\Delta M_{\text{spat}}$ in Fig. 4c does not appear to show a general shrinking or expansion of the convective regions, which would involve systematic mass-flux shifts from the edges of ascent regions toward their interior, or vice versa. This supports the analysis of Hoyos and Webster (2012) for the CMIP3 models, which showed that the area of the dynamical warm pool, defined as the region where net tropospheric diabatic heating is positive, remains constant under warming.

5. Relating mass-flux change to precipitation change

Tropical precipitation change under warming is highly related to changes in convective mass flux but is also influenced by changes in moisture. Therefore, in order to relate the analysis of mass-flux change described in section 4 to rainfall change, we return to the approximation of Eq. (1), $P = Mq$. The Held and Soden (2006) formulation calls for the mass flux from the boundary layer to the free troposphere to be used as $M$, but in practice this is a somewhat idealized concept; it is not possible to obtain a reasonable quantitative estimate of this idealized $M$ from the CMIP5 convective mass-flux data. In the previous sections we have used $M_{\text{int}}$ as a measure of the strength of convection, but $M_{\text{int}}$ is not quantitatively the same as the $M$ of Eq. (1). Therefore, we return to the proxy mass flux $M^*$ of Eq. (2). The use of $M^*$ is appealing because of its simplicity and its direct relationship to $P$ and $q$, but it must first be shown that it is a reasonable proxy for GCM convective mass flux on a grid point by grid point basis; therefore, Eq. (1) is a reasonable assumption when applied locally in the tropics. This is demonstrated in detail in the appendix.

The decomposition of mass-flux change described in section 4 can now be related to changes in precipitation. By construction $P = M^*q$, so the climate change perturbation to $P$, $\Delta P$, is given by

$$\Delta P = \Delta(M^*q)$$

$$= M^* \Delta q + q \Delta M^* + \Delta q M^*,$$

where $M^*$ and $q$ are the climatological values of proxy mass flux and 2-m specific humidity, respectively.

Figure 7 shows $\Delta M^*$ (normalized by global mean temperature change) plotted against $M^*$ for the multimodel mean. The difference $\Delta M^*$ shows very similar behavior.
to $\Delta M_{\text{int}}$ and so can reasonably be decomposed into divergence feedback and spatial components. Values of $\alpha$ were calculated using tropical mean values of $M^*$ and $\Delta M^*$ in the same way as for $M_{\text{int}}$ and are shown in Table 1. The components of $\Delta M^*$ are shown for the multimodel mean in Fig. 8 and are largely similar to the corresponding figures for $M_{\text{int}}$ (Fig. 4), with some minor differences, such as very low values of $\Delta M^*_{\text{div}}$ over the Andes (seen as the lowest values of $\Delta M^*$ in Fig. 7).

Equation (7) can now be broken down further:

$$\Delta P = M^* \Delta q_{\text{CC}} + M^* \Delta q_{\text{RH}} + \Delta q \Delta M^*_{\text{div}} + \Delta q \Delta M^*_{\text{spat}}, (8)$$

where $\Delta q_{\text{CC}}$ is the Clausius–Clapeyron change in 2-m $q$ expected under fixed relative humidity. It is calculated from the local 2-m temperature increase using the August–Roche–Magnus formula (Lawrence 2005), assuming the percentage change in $q_{\text{CC}}$ is equal to the percentage change in saturation vapor pressure under fixed relative humidity. The value $\Delta q_{\text{RH}}$ is the residual $\Delta q - \Delta q_{\text{CC}}$ due to changes in 2-m RH, $\Delta M^*_{\text{div}} = -\alpha M^*$ is the divergence feedback component of $\Delta M^*$, and $\Delta M^*_{\text{spat}} = \Delta M^* - \Delta M^*_{\text{div}}$ is the spatial component of $\Delta M^*$.

The value $\Delta P$ can then be formulated as a combination of different components of tropical precipitation change:

$$\Delta P = \Delta P_T + \Delta P_{\text{RH}} + \Delta P_{\text{div}} + \Delta P_{\text{spat}} + \Delta P_{\text{NL}}, (9)$$

where $\Delta P_T = M^* \Delta q_{\text{CC}}$ is the thermodynamic change in $P$ due to Clausius–Clapeyron–related increases in specific humidity, $\Delta P_{\text{RH}} = M^* \Delta q_{\text{RH}}$ is the change due to near-surface relative humidity changes, $\Delta P_{\text{div}} = q \Delta M^*_{\text{div}}$ is the change due to the divergence feedback on convective mass flux, $\Delta P_{\text{spat}} = q \Delta M^*_{\text{spat}}$ is the change due to spatial shifts in the pattern of mass flux, and $\Delta P_{\text{NL}} = \Delta q \Delta M^*$ is the nonlinear component of precipitation change. These components can easily be calculated and will be analyzed in section 6.

6. Components of tropical precipitation change

a. Analysis of components

Figure 9a shows the multimodel mean pattern of tropical precipitation change between 1971–2000 and 2071–2100 for the RCP8.5 scenario. The rich-get-richer framework (Held and Soden 2006; Chou et al. 2009) of hydrological cycle change, where changes in the magnitude of moisture flux lead to an increase in the gradient of $P - E$, works well for zonal means of $P - E$ (Seager et al. 2010). However, this is not true of the spatial pattern of precipitation change in the tropics, where the spatial correlation between $\Delta P$ and $P$ is 0.2 for
the multimodel mean and a maximum of 0.3 across the CMIP5 models. As specific humidity in the tropics increases approximately as expected from Clausius–Clapeyron, it is curious that this thermodynamic component of precipitation change does not play more of a role in determining the spatial change in precipitation patterns.

Figures 9c and 9d show $\Delta P_T$ and $\Delta P_{\text{div}}$ for the multimodel mean, and a large cancellation between the two components is immediately obvious. The shape of $\Delta P_T = M^* \Delta q_{\text{CC}}$ is mainly determined by $M^*$, which is far more spatially variable than $\Delta q_{\text{CC}}$. Similarly, $\Delta P_{\text{div}} = q \Delta M_{\text{div}}^*$ is mainly shaped by $\Delta M_{\text{div}}^* = -\alpha M^*$ and so is also largely proportional to $M^*$. Figure 9g shows the sum of $\Delta P_T$ and $\Delta P_{\text{div}}$. The cancellation between the two leaves only a relatively small signal over the tropical oceans, with a weak rich-get-richer pattern of larger increases in rainfall in climatological ascent regions. Thus, the divergence feedback on warming acts to oppose the spatial signal of precipitation change due to Clausius–Clapeyron specific humidity increases. Over land the sum of $\Delta P_T$ and $\Delta P_{\text{div}}$ is greater because of the enhanced land warming seen under greenhouse forcing (Dong et al. 2009) leading to large local $\Delta P_T$.

The change $\Delta P_{\text{RH}}$ is shown in Fig. 9e. It is small over the oceans but significant and negative in many tropical land areas, where decreases in relative humidity serve to balance some of the increased rainfall due to $\Delta P_T$. The dominant influence on the multimodel mean pattern of tropical precipitation change is $\Delta P_{\text{spat}}$, shown in Fig. 9b. This spatial component of precipitation change, associated with mechanisms such as SST gradient change that can alter the patterns of convergence and convection, is the main driver of multimodel mean precipitation change in almost all tropical regions, with minor modulation contributed by the other components.

The nonlinear component of precipitation change $\Delta P_{\text{NL}}$ is shown in Fig. 9f. It is relatively small, and in general further opposes the small rich-get-richer signal of $\Delta P_T + \Delta P_{\text{div}}$ with negative precipitation anomalies over climatological ascent regions. This is due to the largest increases in $\Delta q_{\text{CC}}$ and the largest decreases in $\Delta M^*$, both occurring in the ascent regions. There is also
a small positive anomaly in the central equatorial Pacific, associated with the large positive $\Delta M_{\text{spat}}^\text{DM}$ anomaly in this region. The cumulative effect of $\Delta P_{\text{div}}$ and $\Delta P_{\text{NL}}$ is to almost totally cancel out the spatial signal of $\Delta P_T$ over the oceans, allowing the pattern of rainfall change to be largely determined by $\Delta P_{\text{spat}}$.

Figure 9h shows the difference between the multimodel mean precipitation change and the sum of the components on the right-hand side of Eq. (9). The differences are very small throughout the tropics (note the different color scale to that used in the other plots in Fig. 9), demonstrating that the decomposition of Eq. (9) is in this case self-consistent.

One point of interest is that although the thermodynamic component $\Delta P_T$ clearly shows a wet-get-wetter signal, it does not produce a corresponding dry-get-drier pattern. An increase in boundary layer $q$ without circulation change cannot lead to a dry-get-drier response in precipitation in any region, though it may do so in $P - E$. So if the dry-get-drier response does occur in some regions for $P$, then it must be brought about by circulation changes in those regions. Scheff and Frierson (2012) suggest that the observed subtropical drying in CMIP3 models is a shift of the extratropical storm track rather than a dry-get-drier response, so the dry-get-drier mechanism may in fact only apply to $P - E$, not $P$ alone.

Figures 10 and 11 show the zonal mean and equatorial mean changes in the precipitation change components for each CMIP5 model individually. The value $\Delta P_{\text{spat}}$ dominates the variability in the pattern of precipitation change across models, as it does for the multimodel mean, with the magnitude of change being modulated by the remainder of the other components. Over land $\Delta P_{\text{RH}}$ is large in some models, particularly over South America, and can significantly affect the total precipitation change. To illustrate this, Fig. 12 shows the components of equatorial mean precipitation change for HadGEM2-ES, which has large negative $\Delta P_{\text{RH}}$ in some regions, with land areas marked. Over South America the magnitude of $\Delta P_{\text{RH}}$ is larger than that of $\Delta P_{\text{spat}}$ and contributes significantly to the overall negative anomaly, whereas over Africa a smaller negative $\Delta P_{\text{RH}}$ anomaly does not lead to a decrease in total precipitation.

The contribution of each component to the tropical mean precipitation change is shown for each model in Table 2. In contrast to the analysis of spatial patterns, $\Delta P_{\text{spat}}$ is unimportant in the tropical mean, and the total precipitation change is dominated by the results from $\Delta P_{\text{div}}$, $\Delta P_{\text{RH}}$, and $\Delta P_{\text{spat}}$.
change is largely determined by the positive residual of the balance between $\Delta P_T$ and $\Delta P_{\text{div}}$, with a smaller but significant contribution from $\Delta P_{\text{NL}}$.

b. Codependence of the precipitation change components

The decomposition of $\Delta P$ into components is a useful method of gaining insight into the mechanisms that drive precipitation change, but the different components are not completely independent of one another. In this framework, $\Delta P_{\text{div}}$ is determined entirely by the climatological fields and the magnitude of the divergence feedback and so is the most independent of the components. This can be described as the direct effect of the weakening of the tropical circulation on rainfall change. The change $\Delta P_T$ can be affected by changes in $\Delta q_{CC}$ driven by other components, but its shape is largely determined by the climatological mass-flux field and so is relatively independent of the other components.

The value $\Delta P_{\text{spat}}$ can be driven by any of the other components if they lead to shifts in convergence zones. In particular, $\Delta P_{\text{div}}$ and the associated weakening of surface winds is likely to cause SST pattern changes through evaporation and cloud changes, leading to changes in $\Delta P_{\text{spat}}$. These evaporation changes could also potentially affect $\Delta P_{\text{RH}}$. Such mechanisms could

![Fig. 11](image1.png) As in Fig. 10, but for the equatorial mean (5°S–5°N).

![Fig. 12](image2.png) Equatorial mean (5°S–5°N) components of precipitation change (mm day$^{-1}$, normalized by change in global mean $T$) between 1971–2000 and 2071–2100 for HadGEM2-ES. Dashed lines indicate approximate continental boundaries.
be described as the indirect effects of the weakening of the tropical circulation on precipitation change. As the weakening of the tropical circulation is robust across models, whereas other drivers of \( \Delta P_{\text{spat}} \) may not be, these indirect effects of \( \Delta P_{\text{div}} \) may be associated with drivers of \( \Delta P_{\text{spat}} \) that are common to all models such as the enhanced warming of equatorial Pacific SSTs.

Over land \( \Delta P_{\text{spat}} \) and \( \Delta P_{\text{RH}} \) are closely linked, where changes in circulation could lead to changes in moisture influx and relative humidity. Equally, local RH changes could lead to changes in convection over land, altering convection patterns and influencing \( \Delta P_{\text{spat}} \). The magnitude of \( \Delta P_{\text{spat}} \) and \( \Delta P_{\text{RH}} \) appear to be closely coupled over equatorial South America in CMIP5 models (see Fig. 11), but less so over equatorial Africa, and the reasons behind this difference are worthy of further investigation. All the other components are linked to \( \Delta P_{\text{NL}} \), but its variation across models appears to be most affected by large changes in \( \Delta P_{\text{spat}} \).

### 7. Summary and conclusions

The first part of this study examined the dynamical mechanisms that lie behind the patterns of tropical circulation and convective mass-flux change in CMIP5 GCMs under warming. This was then combined with previous knowledge about thermodynamic changes in atmospheric moisture content to construct a simple method of decomposing precipitation change into its constituent parts. Finally, the relative importance of each of these mechanisms in determining the pattern of CMIP5 precipitation change was quantified and the interactions between them examined. Here the main points of interest are summarized and their implications discussed.

#### a. Changes to the tropical circulation and convective mass flux

A robust weakening of the tropical circulation in response to warming occurs for all CMIP5 models under the RCP8.5 scenario. This is consistent with the constraint imposed on tropical circulation change by a greater increase of specific humidity than precipitation in GCMs (Held and Soden 2006). A divergence feedback consistent with the MASC mechanism of Ma et al. (2012) is associated with this weakening, leading to a component of convective mass-flux change under warming that is inversely proportional to the climatological mass-flux field. This is in direct contrast to the convergence feedback that operates in response to localized SST and heating anomalies under natural variability.

The remaining component of mass-flux change is associated with spatial changes in the patterns of low-level convergence and convection, caused by mechanisms such as SST gradient changes, land–sea temperature contrast changes, land surface changes, and local changes in atmospheric dynamics. The largest of these spatial changes are in the central Pacific and are likely to be associated with the equatorially enhanced SST warming seen there (Liu et al. 2005; Xie et al. 2010). In general over the oceans the pattern of spatial mass-flux change is consistent with the hypothesis that SST pattern changes play a dominant role in determining changes in the position of low-level convergence and convection (Xie et al. 2010).

The divergence feedback component of mass-flux change can be thought of as the direct effect of the weakening of the tropical circulation on convective mass flux. Indirect effects of the weakened circulation, such as weakened surface winds causing reduced evaporation and increased SSTs, will also influence the spatial component of mass-flux change.

The framework used here for separation of mass flux into components is perhaps the simplest consistent with the CMIP5 mass-flux data. The concept of a weakening circulation being reflected in a constant fractional decrease in mass flux across the tropics appears to support the idea of a relatively simple, top-down model of a single mechanism such as MASC driving the slowdown of the circulation, rather than a bottom-up combination of various mechanisms acting in different ways in different regions. However, the simple linear relation between \( \Delta M_{\text{int}} \) and climatological \( M_{\text{int}} \) may well need refinement if the divergence feedback associated with the MASC mechanism does not act in a wholly uniform manner across the tropics or if other mechanisms do prove to be important. It is interesting to note that although a divergence rather than a convergence feedback operates at large scales under warming, local convergence feedbacks...
in response to SST pattern change are likely to form part of \( \Delta M_{\text{spat}} \) and to be superimposed on to the large-scale divergence feedback of \( \Delta M_{\text{div}} \).

One mechanism of circulation change not explicitly considered in this study is the direct response of the atmosphere to CO\(_2\) forcing, which acts to suppress convective mass flux due to a stabilization of the troposphere (Andrews et al. 2009; Dong et al. 2009; Chadwick et al. 2013). This will play a part in the weakening of the tropical circulation, though it is not crucial as the circulation weakens even in uniform SST experiments with no greenhouse gas forcing (Ma et al. 2012). As well as stabilizing the atmosphere, the direct response to CO\(_2\) might also contribute to the MASC mechanism by causing more warming in the descent than the ascent regions due to there being less water vapor (WV) and therefore less overlap of WV and CO\(_2\) absorption lines in the subsiding atmosphere. This direct radiative response may manifest itself largely within \( \Delta M_{\text{div}} \).

b. Mechanisms behind the patterns of tropical precipitation change

Precipitation change under warming can be understood as a sum of different components, associated with thermodynamic changes in boundary layer specific humidity, changes in low-level relative humidity, divergence feedback changes in convective mass flux, and spatial changes in mass flux. The rich-get-richer mechanism due to increases in specific humidity and associated changes in moisture transport (Held and Soden 2006; Chou et al. 2009) does not dominate the pattern of precipitation change in the tropics. Instead, there is a large cancellation over the tropical oceans between the spatial patterns of the thermodynamic and divergence feedback components, leaving the pattern of change to be largely determined by the component of precipitation change associated with spatial shifts in convective mass flux.

This spatial component dominates both the pattern of the CMIP5 multimodel mean precipitation change and the intermodel uncertainty in this pattern. Over land the cancellation between \( \Delta P_T \) and \( \Delta P_{\text{div}} \) is less complete because of enhanced land warming (Dong et al. 2009), and the relative humidity component of precipitation change also becomes important. In contrast to the results for spatial patterns, in the tropical mean \( \Delta P_{\text{spat}} \) is negligible, and the observed precipitation increase across models is largely due to the positive remainder of the balance between \( \Delta P_T \) and \( \Delta P_{\text{div}} \).

The spatial cancellation between \( \Delta P_T \) and \( \Delta P_{\text{div}} \) arises because both fields are approximately proportional to the climatological mass-flux field, which, in the case of \( \Delta P_{\text{div}} \), is a consequence of the linearity assumption made in the definition of \( \Delta M_{\text{div}} \) [see Eq. (4)]. This cancellation may prove to be less complete if the divergence feedback associated with the MASC mechanism is found not to be totally linear.

Although these components of precipitation change are interlinked, the decomposition described here does lead to some physical insight into the processes that drive the spatial patterns of rainfall change. It should, however, be noted that care is needed when interpreting the results of this analysis in regions where \( M^* \) differs qualitatively from \( M_{\text{int}} \) (or \( \Delta M^* \) from \( \Delta M_{\text{int}} \)), such as over steep orography in the Andes and Himalayas. The next step would be to investigate further the links between the different components and how these contribute to agreement and uncertainty in precipitation projections across climate models. In particular, the indirect effect of the weakening of the tropical circulation on \( \Delta P_{\text{spat}} \) and the question of how robust this is across models might lead to greater understanding of the patterns of precipitation change under warming.

A recent related result on precipitation change is that of Scheff and Frierson (2012), who found that although \( P - E \) change is consistent with the thermodynamic rich-get-richer mechanism in CMIP3 models, subtropical precipitation declines are not primarily related to thermodynamic changes but are instead due to poleward movement of the midlatitude storm tracks. When this analysis is combined with the results of the current study, it appears that, in general, the extrapolation of the rich-get-richer hypothesis from \( P - E \) to \( P \) alone is not correct.

This study indicates the importance of understanding the range of SST pattern responses to greenhouse gas forcing seen in GCMs, as these are likely to play a major role in the \( \Delta P_{\text{spat}} \) component that dominates the pattern of tropical precipitation change. Possible reasons for the diversity of SST projections across models include model SST biases and differences in parameterizations, such as that of convection. In particular, the variety of vertical heating profiles resulting from different convection schemes (and associated with the wide range of vertical mass-flux profiles seen in Fig. 2) may contribute to this uncertainty. Understanding the connections between GCM SST biases and future SST and precipitation projection patterns may enable constraint of regional rainfall projections by observations of climatological SSTs, such as that proposed by Shiogama et al. (2011), potentially providing more useful information for users of climate projections in tropical regions than is currently available.

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APPENDIX

Relating $M^*$ to $M_{\text{int}}$

Figure 13 shows a scatterplot of $M^*$ versus $M_{\text{int}}$ over the tropical region for the CMIP5 models and the multimodel mean for both the 1971–2000 and 2071–2100 periods. There is a strong correlation between the two quantities (shown in Fig. 13 for 1971–2000), demonstrating that in each model $M^*$ is a reasonable proxy for the convective mass flux in the tropics. Although the two variables have different magnitudes (and units), they behave in a qualitatively very similar manner. Much of the scatter between $M^*$ and $M_{\text{int}}$ exists below the proportionality line, implying that there is precipitation that is not linked to the local convective mass flux. Comparing maps of $M^*$ and $M_{\text{int}}$ (shown for the multimodel mean as the line contours in Figs. 4a and 8a) demonstrates that much of this precipitation occurs over land, particularly on the slopes of steep orography (the Andes and Himalayas), where the precipitation is linked to large-scale ascent.

Figure 13 shows that there is no significant change in the relationship between $M^*$ and $M_{\text{int}}$ in the future climate, demonstrating that the use of $M^*$ to interpret tropical precipitation changes is applicable. Values of the high spatial coefficient of correlation between $\Delta M^*$
and $\Delta M_{\text{int}}$ are shown in Table 1. The weakening of the tropical circulation is also noticeable in Fig. 13, with a shift of both $M^*$ and $M_{\text{int}}$ toward zero for all models. As $M^*$ is calculated with annual means, the subannual covariance between $P$ and $q$ is neglected in Eq. (2). However, for the HadGEM2-ES run, calculating $M^*$ using daily means of $P$ and $q$ gave almost identical results to calculating it using monthly or annual means, so annual means appear to be sufficient.

Reasons for the high degree of correlation between $M^*$ and $M_{\text{int}}$ can be uncovered from Fig. 14. Figures 14a and 14c show the zonal mean moisture divergence ($V \cdot \nabla q$); the solid black line represents the mean convective cloud base and top, and the dashed line represents the schematic level through which $M^*$ is the mass flux. (b),(d) Column-integrated meridional moisture transport ($\overline{\partial q}$). The thin black lines on all plots mark 30°N and 30°S.

FIG. 14. Results from the HadGEM2-ES simulation for the current and future climates. (a),(c) Zonal mean moisture divergence ($V \cdot \nabla q$); the solid black line represents the mean convective cloud base and top, and the dashed line represents the schematic level through which $M^*$ is the mass flux. (b),(d) Column-integrated meridional moisture transport ($\overline{\partial q}$). The thin black lines on all plots mark 30°N and 30°S.

The implication of this is that, on averaged time scales, gridpoint precipitation is well constrained by local boundary layer $q$ and local convective mass flux, meaning that $P$ and $q$ alone can be used to produce a physically meaningful value of mass flux ($M^*$) via Eq. (2). This interpretation of $M^*$ relies on 2-m $q$ being representative of boundary layer $q$ as a whole and, thus, is most valid in well-mixed boundary layers.

One small caveat to this dominance by local moisture ventilation is in the trade wind regions, where moisture ventilated from the subcloud to cloud layer by non-precipitating shallow convection can be transported meridionally and entrained into the deep convection on the ITCZ. This process is responsible for most of the moisture convergence that can be seen above cloud base and the remainder of the scatter on Fig. 13 (as this nonprecipitating shallow convection will be included in $M_{\text{int}}$ but not $M^*$). Taking this within-cloud horizontal transport into account, $M^*$ can be thought of as the local upward transport of mass from the boundary layer through an idealized surface above which moisture convergence is negligible, and therefore moisture is locally precipitated. This surface is generally near but above cloud base and is shown as a schematic in Figs. 14a and 14c. This is a minor extension of the Held and Soden (2006) assumptions behind Eq. (1), allowing it to be applied locally.
It is worth noting that Held and Soden (2006) specify that mass flux due to nonprecipitating shallow convection should be excluded from the calculation of $M$ in Eq. (1). However, it is not clear that this is correct, as moisture transported from the boundary layer to the free troposphere by shallow convection may be precipitated out at a later stage. In any case, although $M_{\text{int}}$ does not exclude shallow convection, it is likely to be dominated by deep convection, due its integration over the depth of the troposphere.

It is noticeable on Figs. 14a and 14c that there is an imbalance in the tropical moisture cycle, with greater convergence of moisture in the extratropics than divergence, implying that the tropics must be exporting some moisture to the extratropics. We calculate this to be $\approx 12\%$ of the tropical surface moisture flux. Figures 14b and 14d show the column integrated meridional moisture flux, showing the regions where moisture is exported from the tropics. This shows that the mid-latitude storm tracks (particularly in the North Atlantic and North Pacific) start just south of 30°N and transport moisture away from the tropics. As discussed by Boutle et al. (2011), the moisture cycle of a midlatitude cyclone is a relatively closed system, with the precipitation balanced by moisture export from the boundary layer due to large-scale ascent. Therefore, this moisture transport appears in neither $M^*$ (because the precipitation falls outside the tropics) nor $M_{\text{int}}$ (because the transport is large scale rather than convective), and the tropics can be considered a closed system with respect to these variables (this was assumed in the calculation of $\alpha$ in section 4).

Therefore, we are confident that Eq. (1) is a good local representation of the relationship between convective mass flux and tropical precipitation and that either $M^*$ or $M_{\text{int}}$ can be used as effective measures of the tropical convective mass flux and changes that occur. The idea here is not that $M^*$ and $M_{\text{int}}$ are identical, but that $M^*$ behaves in a similar way to $M_{\text{int}}$ and the knowledge gained about $\Delta M_{\text{int}}$ in section 4 can reasonably be applied to $\Delta M^*$.

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