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

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Abstract

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

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

1. INTRODUCTION

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

Previous research focused on the effects of the global-mean increase of sea surface temperature (SST) as the zero-order problem, in order to define constraints for climate projections. The most important large-scale response of the circulation is a weakening in the tropical troposphere. Radiative and thermodynamic relations were proposed to diagnostically explain this slow-down (Knutson & Manabe 1995; Allen & Ingram 2002; Stephens & Ellis 2008), and the mean advection of stratification change (MASC) was more recently suggested (Ma et al. 2012) to explain how the atmospheric warming dynamically weakens the tropical circulation. Changes in the lapse rate have also been found to be responsible for the robust poleward expansion of the subsiding branches of the Hadley circulation in observations (Hu &

Fu 2007) and general circulation model simulations (Frierson et al. 2007; Lu et al. 2007; Johanson & Fu 2009). However, these arguments are insufficient to explain why the intensity change of the Hadley cells is not as robust as the Walker cell (Vecchi & Soden 2007a).

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

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

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

The paper consists of three parts: (1) existing and proposed theories; (2) diagnosed mechanisms framed with spatial scales of the changes (e.g., global and regional, ocean and land); (3) future research directions. The theories for precipitation, the Walker, and Hadley circulations (2.1) are summarized in Section 2, followed by their predictions (2.2). Section 3 first introduces the diagnostic slow-down of the tropical circulation and related physical mechanisms (3.1), and then discusses regional precipitation change throughout the tropics (3.2). Section 4 reviews the unique mechanism dominating precipitation change over oceans (4,1) and influencing the structure of the Hadley circulation (4.2). Complicated change of land rainfall (5.2) is summarized in section 5 with introduction of the direct radiative effect of CO_2 (5.1). Potential future research directions are discussed in section 6, followed by a summary in section 7.

2. THEORETICAL FUNDAMENTALS AND PREDICTIONS

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

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

Regions with larger gross moist stability require smaller vertical circulation strength to export a given amount of MSE than do regions of smaller gross moist stability and therefore can be thought of as being more 'stable' in a large-scale sense. Conversely, lower gross moist stability translates to larger upward velocities, and hence less stability, for a given export of MSE. Strong surface low-level wind and moisture convergence over warm waters provides sufficient instability and latent heating to drive intense convection, so that the tropical precipitation bands are associated with sea surface temperature (SST) of at least 26.5-27°C. The thermodynamic coupling between SST and deep convection also plays an important role in shaping rainfall change in global warming.

2.1 Theories for Tropical Tropospheric Overturning Circulations

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

2.1.1 Walker circulation and Bjerknes feedback

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

2.1.2 Hadley circulation and northward displaced ITCZ

On average, the tropics receive more solar radiation than the extratropics, and this energy imbalance is equilibrated by the heat transport via oceanic and atmospheric motion. The Hadley circulation is an important agent for poleward dry energy transport by the tropical atmosphere. This zonal-mean cell is located in both hemispheres and features ascent near the equator and subsidence in the subtropics. The poleward flow in the upper troposphere is turned by the Coriolis force, forming strong westerly subtropical jets at the poleward edges of the Hadley cells, with the help of eddy momentum fluxes. In contrast, the air moves equatorward at the surface, and the Coriolis force results in easterly trade winds.

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

2.1.3 Latitudinal boundary of the Hadley cell

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

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

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

2.2 Future Changes Predicted by the Theories

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

2.2.1 Poleward expansion of the tropics

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

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

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

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

There is accumulating observational evidence confirming the theoretical prediction that the subsidence boundaries of the tropics in both hemispheres are expanding poleward. In a critical assessment, Lucas et al. (2014) identified five methodologies used to define the edge of the tropics in previous studies of tropical expansion: tropopause height frequency, outgoing longwave radiation, total ozone, cloud coverage, temperature, streamfunction, jet stream, and precipitation.

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

2.2.2 Moist static energy constraint for tropical circulation

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

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

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

2.2.3 The "rich-get-richer" theory for regional rainfall change

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

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

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

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

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

3.1 Weakening of the Large-Scale Atmospheric Circulation

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

3.1.1 Observed and simulated weakening of tropical circulation

Since the mid-nineteenth century there has been a weakening trend in the Indo-Pacific zonal sea level pressure gradient, which was suggested by Clarke & Lebedev (1996) as evidence of a weakening Walker circulation. Vecchi et al. (2006) demonstrated that model simulations could accurately reproduce the observed trend in sea level pressure gradient, but only when they included anthropogenic forcing, indicating that the weakening Walker circulation was primarily due to human activities. Recent work has raised more uncertainty about the weakening of the surface winds, with observed global and tropical increases in wind speed during 1987-2006 (Wentz et al. 2007; Ma et al. 2016) and a positive trend in the global evaporation since the late 1970's (Yu 2007). A new dataset, which uses wave height to compensate for errors due to a changing level of wind observations (Tokinaga & Xie 2011), suggests that there has indeed been a Walker circulation slow-down over the last 60 years (Tokinaga et al. 2012), though overall 20th century trends of the surface temperature and pressure in the Pacific remain uncertain across different datasets (Solomon & Newman 2012).

This observational evidence supports model projections of further weakening of the tropical atmospheric circulation during the 21st century (Tanaka et al. 2004). Held & Soden (2006) found a reduction in the amount of convection in CMIP3 (Coupled Model Intercomparison Project phase 3) models, which is also present in the more recent CMIP5 projections (Chadwick et al. 2013). Vecchi & Soden (2007a) further reported a reduction in the frequency of strong updrafts and an increase in the frequency of weak updrafts.

Several complementary hypotheses have been proposed to explain the slowing of the tropical tropospheric circulation, and these are outlined here. This large-scale circulation response to general sea surface temperature (SST) warming can be viewed from a variety of different perspectives, such as the differing rates of change in water vapor, lapse rate, precipitation, and radiative cooling. In fact, these diagnostic results are likely to be all part of the same overall mechanism with unsolved questions, besides the convective depth-related constraint (section 2.2.2).

3.1.2 Thermodynamic relation on convective mass-flux change

Held & Soden (2006) combined the Clausius-Clapeyron equation for the saturation vapor pressure with an approximate equation for precipitation,

$$P = M\,q, \qquad (1)$$

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

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

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

3.1.3 Radiative relation on subsidence change

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

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

In the energy budget for subsidence regions, radiative cooling is mainly balanced by adiabatic warming via large-scale descent. Therefore any change in radiative cooling must be balanced by some combination of lapse-rate change and change in circulation strength, which together determine the magnitude of the change in adiabatic warming. As models predict that the lapse rate decreases more than the relatively weak enhancement in radiative cooling, the vertical motion in descent regions has to weaken to maintain the balance.

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

3.1.4 A dynamical mechanism associated with lapse-rate change

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

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

The MASC effect has an analytical expression and can be diagnosed from general circulation models and applied to a linear baroclinic model, to examine the effect of such geographically uneven heating on atmospheric circulation (Watanabe and Kimoto 2000, 2001). The individual effect of MASC significantly weakens tropical thermal-driven circulations including the Walker and Hadley cells and monsoon winds (Ma & Yu 2014a; Qu & Huang 2016), and reduces the meridional air temperature gradient and hence the tropical wind shear through thermal-wind balance. As a purely dry effect, the MASC is able to offset the enhanced latent heating due to moisture increase, because vertical temperature advection nearly balances diabatic heating in the long term-mean large-scale tropical atmospheric dynamics. In addition, global warming features a pronounced tropical-mean SST warming four times larger than the spatial patterns, resulting in significant atmospheric stabilization and hence MASC effect. In order for the linear baroclinic model to properly reproduce circulation change in a fully coupled model under GHG forcing, this effect has to be explicitly assigned since it is not included by default. In contrast, for internal variability such as El Niño, spatial variations in SST anomalies are 40% greater than the mean warming, so that stratification is less affected and latent heating can be used to drive the linear model effectively without MASC.

Furthermore, since the climatological circulation is relatively similar between models and scenarios, as are the tropical-mean moist adiabats, the magnitude of MASC is also fairly independent of model physics and forcing scenarios. **Figure 1** provides evidence for this argument in the form of an empirical orthogonal function analysis of the MASC forcing among 76 simulations under 3 scenarios, including CMIP3 A1B and CMIP5 Representative Concentration Pathways (RCPs) 6.0 and 8.5, corresponding to medium and high GHG emission scenarios. Dominating 68.3% of the total variance, the first leading mode (**Figure 1*a***) shows significant spatial dependence on climatological pressure velocity, which indicates that the inter-model/scenario variability is primarily in the magnitude of the circulation change as a whole, rather than local differences due to model details and radiative forcing. A high correlation of 0.92 between the first principle component against tropical-mean (40°S-40°N) SST increase suggests that the strength of the MASC effect is mainly dependent on the magnitude of spatially uniform warming in each model (**Figure 1*b***).

3.2 Tropics-Wide Changes in Regional Precipitation

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

3.2.1 What do the observations tell?

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

At very large averaged scales over land there is evidence of wet seasons getting wetter and dry seasons drier (Chou et al. 2013a). However at impacts-relevant regional scales over land, observed rainfall trends do not support a “wet-get-wetter, dry-get-drier” paradigm (Greve et al. 2014; Roderick et al. 2014).

Hegerl et al. (2015) summarized these observed hydrological cycle changes in the context of the existing theories and predictions, e.g., “wet-get-wetter”, “muted” precipitation response and intensified weather extremes. However, they suggested that although observations show robust evidence for theoretically derived and numerically predicted changes, uncertainties from the small signal-to-noise ratio of natural variability, and limitations of short, discontinuous, and inhomogeneous observational datasets pose serious difficulties for determining the anthropogenic contributions.

3.2.2 Decomposition of precipitation change mechanisms

A number of studies have used moisture and energy budgets to examine the regional response of circulation and precipitation change to GHG forcing and to decompose the total response into components associated with various mechanisms. Often the total precipitation change is partitioned into changes associated with thermodynamic (atmospheric moisture increases in response to warming) and dynamical (circulation) processes. Dynamical and thermodynamic changes interact to produce the total pattern of rainfall change, where the thermodynamic component usually represents a “rich-get-richer” term due to moisture increases.

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

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

3.2.3 "Wet-get-wetter" vs. the regional weakening of circulation

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

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

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

4. CHANGES RELATED TO THE SEA SURFACE TEMPERATURE PATTERNS

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

4.1 Circulation and Precipitation Changes over the Oceans

An equatorial peak of precipitation change was frequently illustrated in previous studies. Because the global zonal-mean view makes it difficult to distinguish the rainfall patterns from the purely thermodynamic prediction, it was considered “wet-get-wetter”. However, both the full 2D-pattern and zonal mean of the projected rainfall change are correlated weakly with climatological rainfall in the tropics (Ma & Xie 2013), showing the importance of mechanisms other than the “wet-get-wetter”.

4.1.1 The “warmer-get-wetter” paradigm

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

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

4.1.2 Combined mechanism for tropical oceanic rainfall change

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

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

where $\overline{T}$ is tropical-mean SST warming, T^* means SST patterns, α and β are parameters describing the corresponding strength of the response, and subscripts *weak* and *shift* refer to the circulation slow-down and spatial shift of precipitation, respectively. In percentage form, this approximates the full mechanism controlling the regional precipitation response over the tropical oceans.

The global SST warming contains a uniform increase and spatial patterns. If T^* is zero, Eq. (2) represents the rainfall response in the spatially uniform increase in SST (SUSI). As mentioned in section 2.2.3, the general warming moistens the atmosphere to cause "wet-get-wetter" ($\delta q/q$), and at the same time reduces the tropical circulation ($\delta M_{weak}/M$) to mitigate the moisture effect. This results in a "muted" precipitation response ($\beta\,\overline{T}$). However, even if T^*

is moderate in comparison to $\overline{T}$, it can dominate circulation and precipitation change over the SUSI effect in the coupled climate system, since the factor α = 44% K^{-1} is much larger than β = 2% K^{-1}, according to a linear fit of spatial distribution between T^* and $\delta P/P$ in the CMIP3 ensemble mean (Ma & Xie 2013). The SST patterns shift the atmospheric circulation ($\delta M_{shift}/M$), causing a "warmer-get-wetter" ($\alpha\ T^*$) rainfall response.

As moisture budget analyses (Seager et al. 2010; Chadwick et al. 2013) show, "wet-get-wetter" is the thermodynamic component of regional precipitation change, and circulation slow-down and the SST pattern effects are the dynamical components. Because the circulation weakening contributes a "wet-get-drier" effect, counteracting the "wet-get-wetter" effect, the atmospheric circulation change associated with SST patterns ("warmer-get-wetter") dominates the total rainfall redistribution over the tropical oceans.

4.2 Structural Changes of the Hadley Circulation

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

4.2.1 Interhemispheric energy balance and the ITCZ shift

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

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

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

4.2.2 Inhomogeneous change of the Hadley cell intensity

Section 3.1 discussed the slow-down of the tropical tropospheric circulation; however, robust observational evidence of this slow-down is only found for the Walker cell. In contrast, strengthening trends in the Hadley cell over the last few decades are reported based on prevailing reanalysis datasets, though this could be internal multi-decadal variability rather than a forced trend (Quan et al. 2004; Tanaka et al. 2004; Mitas & Clement 2005, 2006). Future climate model projections suggest stronger future weakening of the Walker than the Hadley circulation (Gastineau et al. 2009; Ma & Xie 2013). Current theories on large-scale circulation change are insufficient to explain these different responses, so that regional effects have to be accounted for (Xie et al. 2015). Eq. (2) suggests that the SST pattern effect turns out to be important for the oceanic rainfall change, with the partial offset between an increase in humidity and a decrease in circulation. For circulation change, humidity is irrelevant so that the circulation weakening is as important as the SST pattern effect. Together, they are the dominant dynamical components of circulation change (section 4.1.2) over the ocean.

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

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

5. DIRECT CO_2 EFFECT AND LAND RAINFALL CHANGE

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

5.1 The Direct CO_2 Radiative Effect

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

Bony et al. (2013) found that by the end of the 21st century, under a high GHG forcing scenario, approximately one quarter to one third of the projected mean tropical circulation change is independent of global mean surface warming, and can thus be attributed to the direct CO_2 effect. Because aqua-planet experiments (all land removed and replaced with ocean) show tropical-mean results that are consistent with simulations that include land, they rejected a major contribution from land warming towards these regional circulation changes. Chadwick et al. (2014) followed up this work and confirmed that the tropical mean circulation change is substantially affected by the direct radiative effect, and is therefore to some extent independent of global mean temperature change. However, regional patterns of rainfall change are dominated by surface warming patterns, including both SST pattern change and land-sea temperature contrast change. They suggested that future regional rainfall changes should be studied primarily with coupled models.

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

Model simulations show differences between global precipitation changes during the 20th and 21st centuries (Thorpe & Andrews 2014), which can be attributed to the changing balance between the influence of SST warming and direct atmospheric absorption. The historical precipitation changes little despite increasing SSTs because large direct effects from CO_2 and black carbon oppose the surface warming-induced precipitation increase, while in future scenarios the importance of these direct effects declines and the SST increase dominates in both ensemble mean and uncertainty, and in both global mean and spatial patterns.

5.2 Circulation and Precipitation Changes over Land

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

5.2.1 Amplified land warming and effect on the circulation

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

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

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

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

Clearly this land temperature adjustment is coupled to evaporation and humidity change, and it is also closely related to circulation change. Enhanced warming over land extends from the surface through the lower-troposphere, and this leads to pressure anomalies and circulation changes (Bayr & Dommenget 2013). Additionally, SST-driven general warming in the free troposphere has the effect of stabilizing the atmosphere over land, and suppressing convection (Joshi et al. 2008; Giannini et al. 2013), reducing rainfall and coupling with evaporation changes. These two effects may be in competition with one another, as the surface warming should drive a sea-breeze-type anomalous circulation bringing air from the oceans to land, whereas suppression of convection and convective heating over land would tend to reduce the flow from ocean to land. It is possible that different circulation changes are seen at different vertical levels. Overall, the net effect of tropical-mean SST warming is to reduce convection and rainfall over land (Giannini et al. 2013; Bony et al. 2013; Chadwick et al. 2014; He et al. 2015).

5.2.2 The direct CO_2 effect-associated land warming

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

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

5.2.3 Relative humidity decreases over land

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

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

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

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

Overall, the circulation and water cycle response over tropical land is likely to consist of a combination of responses to the various mechanisms listed here, with different balances of change emerging in different regions. If these mechanisms can be better understood, it should be possible to reduce the currently substantial inter-model uncertainty in future simulations by evaluating the most important physical processes for any given region between models and observations. Other mechanisms may also be important in some regions, yet their understanding is under-developed and will be discussed in the outlook section 6.3.

6. DIRECTIONS FOR FUTURE RESEARCH

This section identifies promising fields for future research, in order to constrain uncertainties and provide more reliable climate projections. We suggest that fundamental understandings are necessary on the mechanisms shaping the changes of sea surface temperature (SST), intertropical convergence zone (ITCZ), atmospheric circulation, tropical cyclone activity, and land rainfall.

6.1 Combining SST Pattern Change and ITCZ Shifts

It is noteworthy that the extratropical forcing of the ITCZ shift associated with oceanic energy transport has been studied mostly with idealized models (see section 4.2.1). However, it is also important to investigate the projections by comprehensive coupled models for which ITCZ shifts have to be consistent with SST pattern changes and tropical air-sea interactions. As mentioned in section 4.1.1, one major feature of the SST warming is the inter-hemispheric asymmetry, with stronger warming to the north contributing to weakening of the Hadley circulation there. Indeed, Friedman et al. (2013) showed an increasing trend of the south-to-north gradient in the observed SST (< 0.8 K) since 1980 and inferred a northward shift of the ITCZ. They showed that this SST gradient trend is simulated by the CMIP5 (Coupled Model Intercomparison Project phase 5) models and is projected to continue increasing significantly in the 21st century.

Figure 3 uses SST and rainfall as indicators to show the projected change in position of the tropical Pacific ITCZ in the 21st century. Simulated by 19 CMIP5 RCP4.5 (Representative Concentration Pathway 4.5) models, the SST maximum not only "shifts" northward, but also expands toward the equator, and the southward expansion is more significant than the northward expansion. This translates to an ITCZ widening at ~0.5° latitude per century toward both hemispheres, consistent with maximum SST warming on the equator and in the northern subtropics. The precipitation change supports this interpretation, showing equatorward expansion, consistent with more climatological precipitation toward the equator and the above-mentioned energy theories.

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

However, the current consensus separately considers the ITCZ shift and SST patterns, which are likely to be “two sides of the same coin”. Long et al. (2016) attempted to link inter-hemispheric energy balance that drives the ITCZ change to the SST patterns; however, more work is needed to fundamentally understand how the SST patterns emerge. We suggest a synthesized budget analysis combining “both sides” – surface fluxes and ocean transport. Besides short- and long-wave radiation, surface latent and sensible heat fluxes are worth examining, since they involve the influence from winds, relative humidity, and air-sea instability.

6.2 Impact of Circulation Change on Tropical Cyclones

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

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

6.3 Balance of Processes Driving Circulation Change over Land

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

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

7. SUMMARY

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

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

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

SUMMARY POINTS

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

FUTURE ISSUES

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

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LITERATURE CITED

Ackerley D, Dommenget D. 2016. Atmosphere-only GCM simulations with prescribed land surface temperatures. *Geosci. Model Dev. Discuss.* doi:10.5194/gmd-2016-6

Allan RP. 2011. Regime dependent changes in global precipitation. *Clim. Dyn.* 39:827-840

Allan RP, Soden BJ, John VO, Ingram W, Good P. 2010. Current changes in tropical precipitation. *Environ. Res. Lett.* 5:025205. doi:10.1088/1748-9326/5/2/025205

Allen MR, Ingram WJ. 2002. Constraints on future changes in the hydrological cycle. *Nature* 419:224-228

An Z, et al. 2015. Global monsoon dynamics and climate change. *Annu. Rev. Earth Planet. Sci.* 43:29-77

Andrews T, Forster PM, Boucher O, Bellouin N, Jones A. 2010. Precipitation, radiative forcing and global temperature change. *Geophys. Res. Lett.* 37:L14701. doi:10.1029/2010GL043991

Back LE, Bretherton CS. 2006. Geographic variability in the export of moist static energy and vertical motion profiles in the tropical Pacific. *Geophys. Res. Lett.* 33:L17810. doi:10.1029/2006GL026672

Bala G, Caldeira K, Nemani R. 2010. Fast versus slow response in climate change: implications for the global hydrological cycle. *Clim. Dyn.* 35:423-434

Bayr T, Dommenget D. 2013. The tropospheric land-sea warming contrast as the driver of tropical sea level pressure changes. *J. Clim.* 26:1387-1402

Bender MA, et al. 2010. Modeled impact of anthropogenic warming on the frequency of intense Atlantic hurricanes. *Science* 327:454-458

Biasutti M. 2013. Forced Sahel rainfall trends in the CMIP5 archive. *J. Geophys. Res.* 118:1-11

Bjerknes J. 1969. Atmospheric teleconnection from the equatorial Pacific. *Mon. Wea. Rev.* 97:163-172

Bony S, et al. 2013. Robust direct effect of carbon dioxide on tropical circulation and regional precipitation. *Nat. Geosci.* 6:447-451

Bretherton CS, Sobel AH. 2002. A simple model of a convectively-coupled Walker circulation using the weak temperature gradient approximation. *J. Clim.* 15:2907-2920

Bruyère CL, Holland GJ, Towler E. 2012. Investigating the use of a genesis potential index for tropical cyclones in the North Atlantic basin. *J. Clim.* 25:8611-8626

Byrne MP, O’Gorman PA. 2013. Link between land-ocean warming contrast and surface relative humidities in simulations with coupled climate models. *Geophys. Res. Lett.* 40:5223-5227. doi:10.1002/grl.50971

Byrne MP, O’Gorman PA. 2015. The response of precipitation minus evapotranspiration to climate warming: Why the “wet-get-wetter, dry-get-drier” scaling does not hold over land. *J. Clim.* 28:8078-8092

Cao L, Bala G, Caldeira K. 2012. Climate response to changes in atmospheric carbon dioxide and solar irradiance on the time scale of days to weeks. *Environ. Res. Lett.* 7:034015

Camargo SJ, Emanuel KA, Sobel AH. 2007. Use of a genesis potential index to diagnose ENSO effects on tropical cyclone genesis. *J. Clim.* 20:4819-4834

Chadwick R, Boutle I, Martin G. 2013. Spatial patterns of precipitation change in CMIP5: Why the rich do not get richer in the tropics. *J. Clim.* 26:3803-3822

Chadwick R, Good P, Andrews T, Martin G. 2014. Surface warming patterns drive tropical rainfall pattern responses to CO2 forcing on all timescales. *Geophys. Res. Lett.* 41:610-615. doi:10.1002/2013GL058504

Chadwick R. 2016. Which aspects of CO_2 forcing and SST warming cause most uncertainty in projections of tropical rainfall change over land and ocean? *J. Clim.* doi:10.1175/JCLI-D-15-0777.1

Chen G, Held IM. 2007. Phase speed spectra and the recent poleward shift of Southern Hemisphere surface westerlies. *Geophys. Res. Lett.* 34:L21805. doi:10.1029/2007GL031200

Chen G, Lu J, Frierson DMW. 2008. Phase speed spectra and the latitude of surface westerlies: Interannual variability and the global warming trend. *J. Clim.* 21:5942-5959. doi:10.1175/2008JCLI2306.1

Chou C, Neelin JD. 2004. Mechanisms of global warming impacts on regional tropical precipitation. *J. Clim.* 17:2688-2701

Chou C, Chen C-A. 2010. Depth of convection and the weakening of the tropical circulation in global warming. *J. Clim.* 23:3019-3030

Chou C, Chiang JCH, Lan C-W, Chung C-H, Liao Y-C, Lee C-J. 2013a. Increase in the range between wet and dry season precipitation. *Nat. Geosci.* doi:10.1038/NGEO1744

Chou C, Neelin JD, Chen C-A, Tu J-Y. 2009. Evaluating the “rich-get-richer” mechanism in tropical precipitation change under global warming. *J. Clim.* 22:1982-2005

Chou C, Wu T-C, Tan P-H. 2013b. Changes in gross moist stability in the tropics under global warming. *Clim. Dyn.* 41:2481-2496. doi:10.1007/s00382-013-1703-2

Clarke AJ, Lebedev A. 1996. Long-term change in the equatorial Placific trade winds. *J. Clim.* 9:1020-1029

Compo GP, Sardeshmukh PD. 2009. Oceanic influences on recent continental warming. *Clim. Dyn.* 32:333-342

Curry R, Dickson B, Yashayaev I. 2003. A change in the fresh-water balance of the Atlantic Ocean over the past four decades. *Nature* 426:826-829

DeMaria M. 1996. The effect of vertical shear on tropical cyclone intensity change. *J. Atmos. Sci.* 53:2076-2087

Deser C, Phillips AS. 2009. Atmospheric circulation trends, 1950-2000: The relative roles of sea surface temperature forcing and direct atmospheric radiative forcing. *J. Clim.* 22:396-413

Dong B, Gregory JM, Sutton RT. 2009. Understanding land-sea warming contrast in response to increasing greenhouse gases. Part I: Transient adjustment. *J. Clim*. 22:3079-3097

Durack PJ, Wijffels SE. 2010. Fifty-year trends in global ocean salinities and their relationship to broad-scale warming. *J. Clim.* 23:4342-4362

Durack PJ, Wijffels SE, Matear RJ. 2012. Ocean salinities reveal strong global water cycle intensification during 1950 to 2000. *Science* 336(6080):455-458. doi:10.1126/science.1212222

Emanuel KA. 1999. Thermodynamic control of hurricane intensity. *Nature* 401:665-669

Fasullo J. 2012. A mechanism for land-ocean contrasts in global monsoon trends in a warming climate. *Clim. Dyn.* 39:1137-1147

Friedman AR, Hwang Y-T, Chiang JCH, Frierson DMW. 2013. Interhemispheric temperature asymmetry over the twentieth century and in future projections. *J. Clim.* 26:5419-5433

Frierson DMW, Hwang Y-T. 2012. Extratropical influence on ITCZ shifts in slab ocean simulations of global warming. *J. Clim.* 25:720-733

Frierson DMW, Lu J, Chen G. 2007. Width of the Hadley cell in simple and comprehensive general circulation models. *Geophys. Res. Lett.* 34:L18804. doi:10.1029/2007GL031115

Fu Q, Johanson CM, Wallace JM, Reichler T. 2006. Enhanced mid-latitude tropospheric warming in satellite measurements. *Science* 312:1179

Fučkar NS, Xie S-P, Farneti R, Maroon EA, Frierson DMW. 2013. Influence of the extratropical ocean circulation on the intertropical convergence zone in an idealized coupled general circulation model. *J. Clim.* 26:4612-4629

Gastineau G, Li L, Le Treut H. 2009. The Hadley and Walker circulation changes in global warming conditions described by idealized atmospheric simulations. *J. Clim.* 22:3993-4013

Giannini A. 2010. Mechanisms of climate change in the semiarid African Sahel: The local view. *J. Clim.* 23:743-756

Giannini A, Salack S, Lodoun T, Ali A, Gaye AT, Ndiaye O. 2013. A unifying view of climate change in the Sahel linking intra-seasonal, interannual and longer time scales. *Environ. Res. Lett.* 8:024010

Greve P, et al. 2014. Global assessment of trends in wetting and drying over land. *Nat. Geosci.* doi:10.1038/NGEO2247

Hartmann DL, Larson K. 2002. An important constraint on tropical cloud - climate feedback. *Geophys. Res. Lett.* 29:1951. doi:10.1029/2002GL015835

He J, Soden BJ, Kirtman B. 2014. The robustness of the atmospheric circulation and precipitation response to future anthropogenic surface warming. *Geophys. Res. Lett.* 41:2614-2622. doi:10.1002/2014GL059435

He J, Soden BJ. 2015. Anthropogenic weakening of the tropical circulation: The relative roles of direct CO_2 forcing and sea surface temperature change. *J. Clim*. doi:10.1175/JCLI-D-15-0205.1

Hegerl GC, et al. 2015. Challenges in quantifying changes in the global water cycle. *Bull. Amer. Meteor. Soc.* 96:1097-1115

Held IM, Hou AY. 1980. Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *J. Atm. Sci.* 37:515-533

Held IM, Soden BJ. 2000. Water vapor feedback and global warming. *Ann. Rev. Energy Environ. 25*:441-75

Held IM, Soden BJ. 2006. Robust responses of the hydrological cycle to global warming. *J. Clim.* 19:5686-5699

Held IM, Delworth TL, Lu J, Findell KL, Knutson TR. 2005. Simulation of Sahel drought in the 20th and 21st centuries. *Proc. Natl. Acad. Sci.* 102:17891-17896

Hu Y, Fu Q. 2007. Observed poleward expansion of the Hadley circulation since 1979. *Atmos. Chem. Phys.* 7:5229-5236. doi:10.5194/acp-7-5229-2007

Ingram W. 2010. A very simple model for the water vapour feedback on climate change. *Q. J. R. Meteorol. Soc.* 136:30-40

Johanson CM, Fu Q. 2009. Hadley cell widening: Model simulations versus observations. *J. Clim.* 22:2713-2725

Johnson NC, Xie S-P. 2010. Changes in the sea surface temperature threshold for tropical convection. *Nat. Geosci.* 3:842-845. doi:10.1038/ngeo1008

Joshi MM, Gregory JM, Webb MJ, Sexton DMH, Johns TC. 2008. Mechanisms for the land/sea warming contrast exhibited by simulations of climate change. *Clim. Dyn.* 30:455-465

Kamae Y, Watanabe M, Ogura T, Yoshimori M, Shiogama H. 2015. Rapid adjustments of cloud and hydrological cycle to increasing CO_2: A review. *Curr. Clim. Change Rep.* 1:103-113. doi:10.1007/s40641-015-0007-5

Kang SM, Frierson DMW, Held IM. 2009. The tropical response to extratropical thermal forcing in an idealized GCM: The importance of radiative feedbacks and convective parameterization. *J. Atmos. Sci.* 66:2812-2827

Kang SM, Seager R, Frierson DMW, Liu X. 2015. Croll revisited: Why is the northern hemisphere warmer than the southern hemisphere? *Clim. Dyn.* 44:1457-1472

Kidston J, et al. 2015. Stratospheric influence on tropospheric jet streams, storm tracks and surface weather. *Nat. Geosci.* 8:433-440. doi:10.1038/ngeo2424

Knutson TR, Manabe S. 1995. Time-mean response over the tropical Pacific to increased CO_2 in a coupled ocean-atmosphere model. *J. Clim.* 8:2181-2199

Knutson TR, Sirutis JJ, Garner ST, Vecchi GA, Held IM. 2008. Simulated reduction in Atlantic hurricane frequency under twenty-first-century warming conditions. *Nat. Geosci.* 1:359-364

Lambert FH, Webb MJ. 2008. Dependency of global mean precipitation on surface temperature. *Geophys. Res. Lett.* 35:L16706. doi:10.1029/2008GL034838

Lambert FH, Webb MJ, Joshi MM. 2011. The relationship between land-ocean surface temperature contrast and radiative forcing. *J. Clim.* 24:3239-3256

Lau K-M, Wu H-T. 2007. Detecting trends in tropical rainfall characteristics, 1979-2003. *Int. J. Climatol.* 27:979-988

Le Hir G, et al. 2009. The Snowball Earth aftermath: Exploring the limits of continental weathering processes. *Earth Plan. Sci. Lett.* 277:453-463

Lee S-S, Tao W-K, Jung C-H. 2014. Aerosol effects on instability, circulations, clouds, and precipitation. *Adv. Meteorol.* 2014:683950. doi/10.1155/2014/683950

Liu Z, Vavrus S, He F, Wen N, Zhong Y. 2005. Rethinking tropical ocean response to global warming: The enhanced equatorial warming. *J. Clim.* 18:4684-4700

Long S-M, Xie S-P, Liu W. 2016. Uncertainty in tropical rainfall projections: Atmospheric circulation effect and the ocean coupling. *J Clim.* 29:2671-2687

Lorenz DJ, DeWeaver ET. 2007. Tropopause height and zonal wind response to global warming in the IPCC scenario integrations. *J. Geophys. Res.* 112:D10119. doi:10.1029/2006JD008087

Lu J, Zhao B. 2012. The role of oceanic feedback in the climate response to doubling CO_2. *J. Clim.* 25:7544-7563

Lu J, Vecchi GA, Reichler T. 2007. Expansion of the Hadley cell under global warming. *Geophys. Res. Lett.* 34:L06805. doi:10.1029/2006GL028443

Lucas C, Timbal B, Nguyen H. 2014. The expanding tropics: A critical assessment of the observational and modeling studies. *WIREs Clim. Change* 5:89-112. doi:10.1002/wcc.251

Ma J, Xie S-P. 2013. Regional patterns of sea surface temperature change: A source of uncertainty in future projections of precipitation and atmospheric circulation. *J. Clim.* 26:2482-2501. doi:10.1175/JCLI-D-12-00283.1

Ma J, Yu J-Y. 2014a. Paradox in South Asian summer monsoon circulation change: Lower tropospheric strengthening and upper tropospheric weakening. *Geophys. Res. Lett.* 41:2934-2940. doi:10.1002/2014GL059891

Ma J, Yu J-Y. 2014b. Linking centennial surface warming patterns in the equatorial Pacific to the relative strengths of the Walker and Hadley circulations. *J. Atmos. Sci.* 71:3454-3464

Ma J, Foltz GR, Soden BJ, Huang G, He J, Dong C. 2016. Will surface winds weaken in response to global warming? *Env. Res. Lett.* 11:124012

Ma J, Xie S-P, Kosaka Y. 2012. Mechanisms for tropical tropospheric circulation change in response to global warming. *J. Clim.* 25:2979-2994

Meehl GA, et al. 2007. The WCRP CMIP3 multimodel dataset: A new era in climate change research. *Bull. Amer. Meteor. Soc.* 88:1383-1394

Mitas CM, Clement A. 2005. Has the Hadley cell been strengthening in recent decades? *Geophys. Res. Lett.* 32:L03809. doi:10.1029/2004GL021765

Mitas CM, Clement A. 2006. Recent behavior of the Hadley cell and tropical thermodynamics in climate models and reanalyses. *Geophys. Res. Lett.* 33:L01810. doi:10.1029/2005GL024406

Muller CJ, O'Gorman PA. 2011. An energetic perspective on the regional response of precipitation to climate change. *Nat. Clim. Change* 1:266–271. doi:10.1038/nclimate1169

Neelin JD, Zeng N. 2000. A quasi-equilibrium tropical circulation model---formulation. *J. Atmos. Sci.* 57:1741-1766

Neelin JD, Chou C, Su H. 2003. Tropical drought regions in global warming and El Niño teleconnections. *Geophys. Res. Lett.* 30:2275. doi:10.1029/2003GLO018625

O’Gorman PA, Muller C. 2010. How closely do changes in surface and column water vapor follow Clausius-Clapeyron scaling in climate change simulations? *Environ. Res. Lett.* 5:025207. doi:10.1088/1748-9326/5/2/025207

O'Gorman PA, Singh MS. 2013. Vertical structure of warming consistent with an upward shift in the middle and upper troposphere. *Geophys. Res. Lett.* 40:1838-1842

O'Gorman PA, Allan RP, Byrne MP, Previdi M. 2012. Energetic constraints on precipitation under climate change. *Surv. Geophys.* 33:585-608

Pierrehumbert RT. 1995. Thermostats, radiator fins, and the local runaway greenhouse. *J. Atmos. Sci.* 52:1784-1806

Pierrehumbert RT. 1999. Subtropical water vapor as a mediator of rapid global climate change, in Mechanisms of global change at millennial time scales, Clark PU, Webb RS, Keigwin LD eds. *Geophys. Monog. Series* 112:394 pp, American Geophysical Union, Washington, D.C.

Pierrehumbert RT. 2002. The hydrologic cycle in deep time climate problems. *Nature* 419:191-198

Pierrehumbert RT. 2010. Principles of planetary climate: Chapter 6. Cambridge University Press, 652pp

Pierrehumbert RT, Roca R. 1998. Evidence for control of Atlantic subtropical humidity by large scale advection. *Geophys. Res. Lett.* **25:**4537-4540

Qu X, Huang G. 2016. The global warming-induced South Asian High change and its uncertainty. *J. Clim.* doi:10.1175/JCLI-D-15-0638.1

Quan X-W, Diaz HF, Hoerling MP. 2004. Change in the tropical Hadley cell since 1950, in The Hadley circulation: Past, present, and future. Edited by Diaz HF, and Bradley RS, Cambridge Univ. Press, New York

Richardson T, Forster P, Andrews T, Parker D. 2016. Understanding the Rapid Precipitation Response to CO_2 and Aerosol Forcing on a Regional Scale. *J. Clim.* doi:10.1175/JCLI-D-15-0174.1

Richter I, Xie S-P. 2008. The muted precipitation increase in global warming simulations: A surface evaporation perspective. *J. Geophys. Res.-Atmos.* 113:D24118. doi:10.1029/2008JD010561

Roderick ML, Sun F, Lim WH, Farquhar GD. 2014. A general framework for understanding the response of the water cycle to global warming over land and ocean. *Hydrol. Earth Syst. Sci.* 18:1575-1589

Rodwell MJ, Hoskins BJ. 1996. Monsoons and the dynamics of deserts. *Quart. J. Roy. Meteor. Soc.* 122:1385-1404. doi:10.1002/qj.49712253408

Rowell D, Jones R. 2006. Causes and uncertainty of future summer drying over Europe. *Clim. Dyn.* 27:281-299

Scheff J, Frierson D. 2012. Twenty-first-century multimodel subtropical precipitation declines are mostly midlatitude shifts. *J. Clim.* 25:4330-4347

Schneider T, O'Gorman PA, Levine XJ. 2010. Water vapor and the dynamics of climate changes. *Rev. Geophys.* 48:RG3001. doi:10.1029/2009RG000302

Seager R, et al. 2007. Model projections of an imminent transition to a more arid climate in southwestern North America. *Science* 316:1181-1184

Seager R, Naik N, Vecchi GA. 2010. Thermodynamic and dynamic mechanisms for large-scale changes in the hydrological cycle in response to global warming. *J. Clim.* 23:4651-4668

Seidel DJ, Fu Q, Randel WJ, Reichler TJ. 2008. Widening of the tropical belt in a changing climate. *Nat. Geosci.* 1:21-24

Seo J, Kang SM, Frierson DMW. 2014. Sensitivity of intertropical convergence zone movement to the latitudinal position of thermal forcing. *J. Clim.* 27:3035-3042

Simmons A, Willett K, Jones P, Thorne P, Dee D. 2010. Low-frequency variations in surface atmospheric humidity, temperature and precipitation: Inferences from reanalyses and monthly gridded observational datasets. *J. Geophys. Res.* 115:1-21. doi:0.1029/2009JD012442

Singh MS, O'Gorman PA. 2013. Influence of entrainment on the thermal stratification in simulations of radiative-convective equilibrium. *Geophys. Res. Lett.* 40:1-6. doi:10.1002/grl.50796

Sobel AH, Camargo SJ. 2011. Projected future changes in tropical summer climate. *J. Clim.* 24:473-487

Sobel AH, Nilsson J, Polvani LM. 2001. The weak temperature gradient approximation and balanced tropical moisture waves. *J. Atmos. Sci.* 58:3650-3665

Solomon A, Newman M. 2012. Reconciling disparate twentieth-century Indo-Pacific ocean temperature trends in the instrumental record. *Nat. Clim. Change* 2:691-699

Stephens GL, Ellis TD. 2008. Controls of global-mean precipitation increases in global warming GCM experiments. *J. Clim.* 21:6141-6155.

Sun Y, Solomon S, Dai A, Portmann RW. 2007. How often will it rain? *J. Clim*. 20:4081-4818

Sugi M, Yoshimura J. 2003. A mechanism of tropical precipitation change due to CO_2 increase. *J. Clim.* 17:238-243

Sutton RT, Dong B, Gregory JM. 2007. Land/sea warming ratio in response to climate change: IPCC AR4 model results and comparison with observations. *Geophys. Res. Lett.* 34:L02701. doi:10.1029/2006GL028164

Tanaka HL, Ishizaki N, Kitoh A. 2004. Trend and interannual variability of Walker, monsoon and Hadley circulations defined by velocity potential in the upper troposphere. *Tellus* 56A:250-269

Taylor KE, Stouffer RJ, Meehl GA. 2012. An Overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.* 93:485-498. doi:10.1175/BAMS-D-11-00094.1

Thorpe L, Andrews T. 2014. The physical drivers of historical and 21st century global precipitation changes. *Environ. Res. Lett.* 9:064024. doi:10.1088/1748-9326/9/6/064024

Tokinaga H, Xie S-P. 2011. Weakening of the equatorial Atlantic cold tongue over the past six decades. *Nat. Geosci.* 4:222-226

Tokinaga H, Xie S-P, Timmermann A, McGregor S, Ogata T, Kubota H, Okumura YM. 2012. Regional patterns of tropical Indo-Pacific climate change: Evidence of the Walker Circulation weakening. *J. Clim.* 25:1689-1710

Vecchi GA, Soden BJ. 2007a. Global warming and the weakening of the tropical circulation. *J. Clim.* 20:4316-4340

Vecchi GA, Soden BJ. 2007b. Effect of remote sea surface temperature change on tropical cyclone potential intensity. *Nature* 450:1066-1070

Vecchi GA, Soden BJ. 2007c. Increased tropical Atlantic wind shear in model projections of global warming. *Geophys. Res. Lett.* 34:L08702. doi:10.1029/2006GL028905

Vecchi GA, Soden BJ, Wittenberg AT, Held IM, Leetmaa A, Harrison MJ. 2006. Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature* 441:73-76. doi:10.1038/nature04744

Vecchi, GA, Swanson KL, Soden BJ. 2008. Whither hurricane activity? *Science* 322:687-689

Véspoli de Carvalho LM, Jones C. Edited 2016. The monsoons and climate change. Springer, New York. doi:10.1007/978-3-319-21650-8

Voigt A, Shaw TA. 2015. Circulation response to warming shaped by radiative changes of clouds and water vapor. *Nat. Geosci.* 8:102-106. doi:10.1038/ngeo2345

Wang C, Lee S-K. 2008. Global warming and United States landfalling hurricanes. *Geophys. Res. Lett.* 35:L02708. doi:10.1029/2007GL032396

Wang C, Lee S-K, Enfield DB. 2008. Atlantic warm pool acting as a link between Atlantic multidecadal oscillation and Atlantic tropical cyclone activity. *Geochem. Geophys. Geosyst.* 9:Q05V03. doi:10.1029/2007GC001809

Wang Y, JH Jiang, Su H. 2015. Atmospheric responses to the redistribution of anthropogenic aerosols. *J. Geophys. Res.-Atmos.* 120:9625-9641. doi:10.1002/2015JD023665

Watanabe M, Kimoto M. 2000. Atmosphere-ocean thermal coupling in the North Atlantic: A positive feedback. *Quart. J. R. Met. Soc.* 126:3343-3369

Watanabe M, Kimoto M. 2001. Corrigendum. *Quart. J. R. Met. Soc.* 127:733-734

Webb MJ, Andrews T, Bodas-Salcedo A, Bony S, Bretherton CS, Chadwick R, et al. 2017. The Cloud Feedback Model Intercomparison Project (CFMIP) contribution to CMIP6. *Geosci. Model Dev.* 10:359-384

Wentz FJ, Ricciardulli L, Hilburn K, Mears C. 2007. How much more rain will global warming bring? *Science* 317:233-235

Williams IN, Pierrehumbert RT, Huber M. 2009. Global warming, convective threshold and false thermostats. *Geophys. Res. Lett.* 36:272-277

Xie S-P, Philander SGH. 1994. A coupled ocean-atmosphere model of relevance to the ITCZ in the eastern Pacific. *Tellus* 46A:340-350

Xie S-P, Deser C, Vecchi GA, Ma J, Teng H, Wittenberg AT. 2010. Global warming pattern formation: Sea surface temperature and rainfall. *J. Clim.* 23:966-986

Xie S-P, et al. 2015. Towards predictive understanding of regional climate change. *Nat. Clim. Change* 5:921-930

Yang F, Kumar A, Schlesinger ME, Wang W. 2003. Intensity of hydrological cycles in warmer climates. *J. Clim.* 16:2419-2423

Yu L. 2007. Global variations in oceanic evaporation (1958-2005): The role of the changing wind speed. *J. Clim.* 20:5376-5390

Zelinka MD, Hartmann DL. 2011. The observed sensitivity of high clouds to mean surface temperature anomalies in the tropics. *J. Geophys. Res.* 116:D23103. doi:10.1029/2011JD016459

Zeng N, Neelin JD, Chou C. 2000. A quasi-equilibrium tropical circulation model---implementation and simulation. *J. Atmos. Sci.* 57:1767-1796

Zhang X, et al. 2007. Detection of human influence on twentieth-century precipitation trends. *Nature* 448:461-465. doi:10.1038/nature06025

Zhao M, Held IM. 2012. TC-permitting GCM simulations of hurricane frequency response to sea surface temperature anomalies projected for the late 21st Century. *J. Clim.* 25:2995-3009

Zhou YP, Xu KM, Sud YC, Betts AK. 2011. Recent trends of the tropical hydrological cycle inferred from Global Precipitation Climatology Project and International Satellite Cloud Climatology Project data. *J. Geophys. Res.* 116:D09101. doi:10.1029/2010JD015197

ACRONYMS AND DEFINITIONS

Bjerknes feedback: Ocean-atmosphere positive feedback shaping equatorial climate variability/change, e.g., trade wind relaxation causes flattened thermocline, decreased SST gradients and hence further wind weakening;

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

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

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

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

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

GHG: Greenhouse gas, a gas in the atmosphere absorbing, emitting infrared radiation, and warming Earth, e.g., H_2O, CO_2, CH_4, N_2O, O_3;

Hadley circulation (Hadley cell): A large-scale tropical circulation, with air rising near the equator and descending in the subtropics, and flowing equatorward near the surface and poleward near the tropopause;

ITCZ: Intertropical convergence zone, an area encircling Earth where the northeast and southeast trade winds converge within approximately ±20° of the Equator;

MASC: Mean advection of stratification change, a dynamical mechanism that has been proposed to explain the weakening of the tropical tropospheric circulation in climate change, due to tropics-wide warming increasing with height and hence stabilizing the atmosphere;

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

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

Planetary boundary layer: The lowest part of the atmosphere, directly influenced by the surface, with turbulent wind, temperature, moisture, and strong vertical mixing;

RCPs: Representative Concentration Pathways, four GHG concentration (not emissions) trajectories adopted by the IPCC for its fifth Assessment Report in 2014;

RH: Relative humidity, the ratio of the partial pressure of water vapor to the equilibrium vapor pressure at a given temperature;

SRES A1B: Special Report on Emissions Scenarios A1B, GHG emissions peaking in the mid-21st century, balancing across old and new energy sources;

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

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

Walker circulation (Walker cell): A thermally driven equatorial zonal and vertical circulation, e.g., rising above the western Pacific, near the maritime Asian continent, and sinking over the eastern Pacific;

Wind-evaporation-SST feedback: An ocean-atmosphere interaction process shifting the ITCZ north of the equator, involving the Coriolis force and evaporation adjustment.

SIDEBAR

(Sections 1, between Sections 3.1.2 and 3.1.3)
Thermodynamic and radiative relations for tropical circulation: The global-mean water vapor and vertical gradient of air temperature increase at ~7% K^{-1} of surface warming, but the fractional changes in global mean precipitation and net long-wave radiative cooling are only 1-2% K^{-1} based on model projections. This contrast indicates a weakening of the circulation at a rate of ~5% K^{-1} in both convective and subsidence regions.

(Sections 1, 3.1.4, 3.2.3, 6.1, 6.2)
MASC: Mean advection of stratification change, a dynamical mechanism proposed to explain the weakening of the tropospheric circulation throughout the tropics in climate change. Following moist adiabat, tropics-wide warming increases with height and stabilizes the atmosphere. This leads to relative cooling of the air column in ascending regions due to anomalous cold advection of low-level air, and relative warming in subsidence regions due to warm advection. This reduces air temperature and pressure gradients between ascent and descent regions, opposes the climatological circulation, and slows it down as an adiabatic forcing.

(Sections 1, 2.2.3, 3.2, 4.1)
Wet-get-wetter: In the (hypothetical) absence of circulation change, increased atmospheric water vapor implies an increased moisture transport from dry to wet regions, and hence an increased precipitation gradient, i.e., rainfall increases in the core of existing rainy regions, and decreases in current dry areas and at convective margins.

(Sections 1, 4.1, 4.2.2, 6)
Warmer-get-wetter: A general warming of the tropical troposphere raises the sea surface temperature (SST) threshold for tropical convection, so that convective precipitation increases where SST warming exceeds the tropical mean and decreases where relatively weak warming exists. This SST pattern effect is achieved by adjusting the atmospheric circulation and corresponding surface divergence field, including two outstanding modes: an equatorial peak anchoring a local precipitation increase, and a meridional dipole with increased rainfall and weakened trade winds over the warmer hemisphere.

(Sections 2.1.1, 4.1.1, 6.1)

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

(Sections 2.1.2, 4.1.1, 6.1)

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

(Section 5.2.1)

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

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

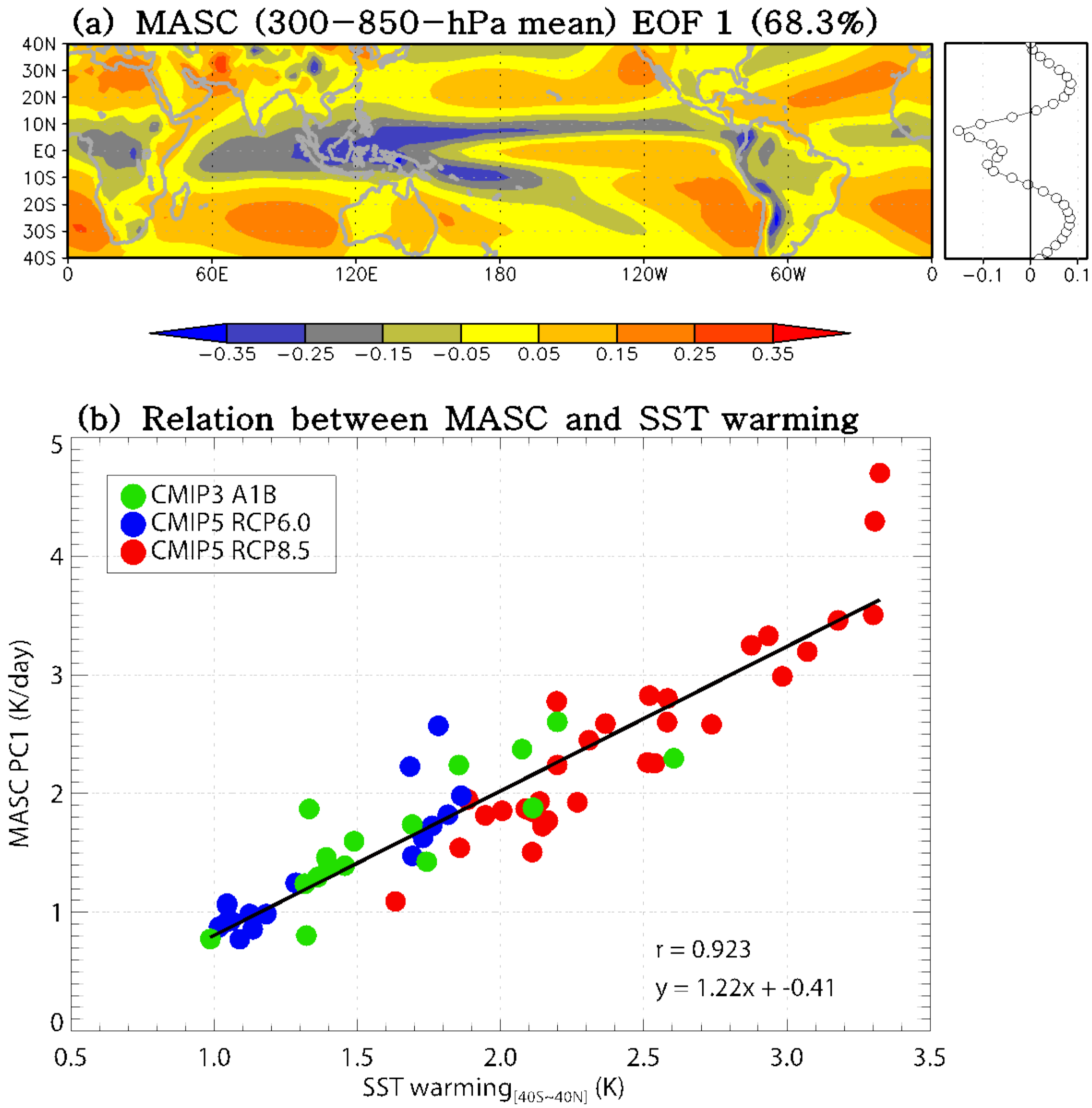

Figure 1 First leading mode of the inter-model empirical orthogonal function analysis for the vertical-averaged MASC forcing calculated with 76 simulations for 3 scenarios: SRES A1B (22) of CMIP3, RCPs 6.0 (23) and 8.5 (30) of CMIP5. Spatial mode 1 (*a*) shows its pattern-dependence only on climatological vertical velocity, and a scatter between tropical-mean SST warming and the first principle component (*b*) indicates its magnitude-dependence only on the general warming extent.

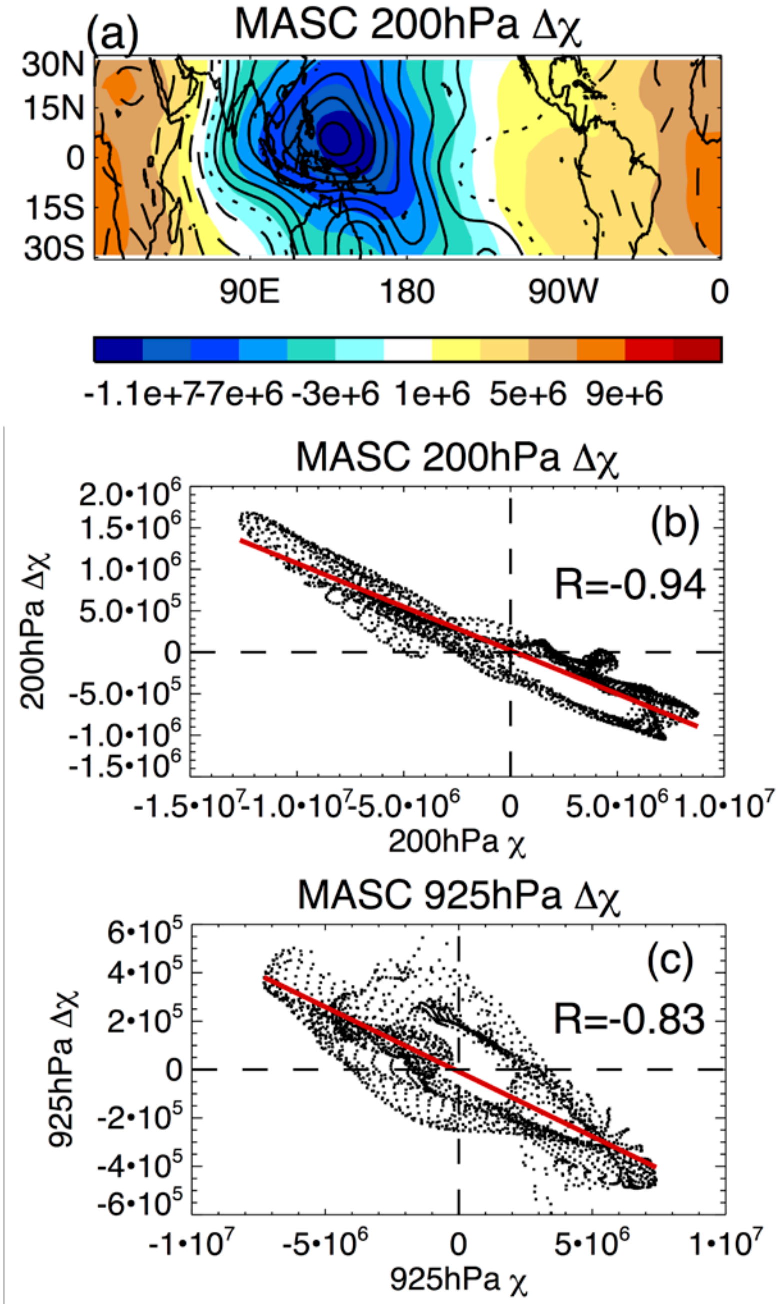

Figure 2 Local weakening of circulation induced by MASC forcing (**Figure 1*a***) in a linear baroclinic model. Represented with velocity potential, (*a*) shows the horizontal distribution of change (color) over climatology (contours) at the 200 hPa level and scatterplots indicate their negative linear relationship at (*b*) 200 hPa and (*c*) 925 hPa.

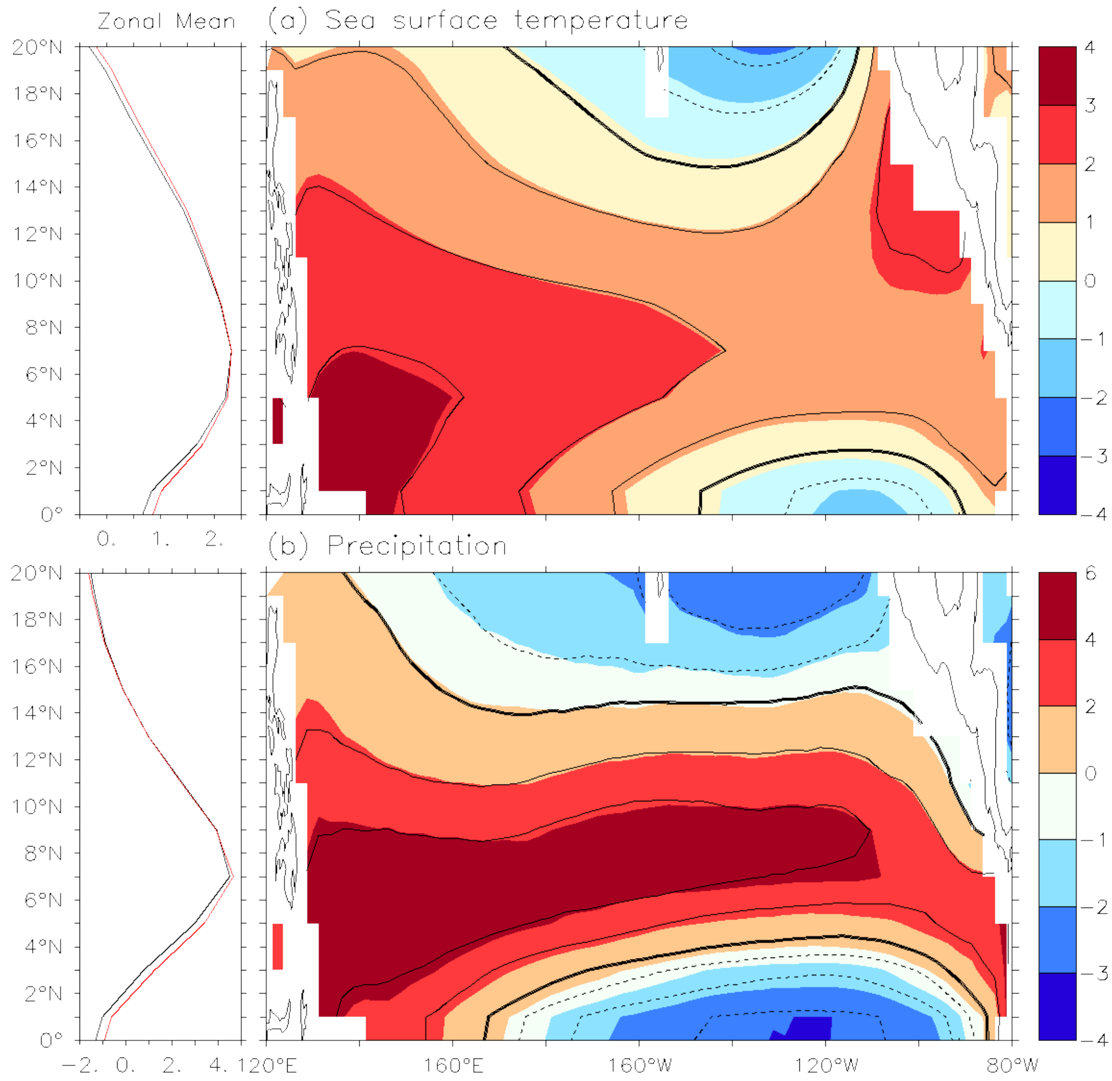

Figure 3 Ensemble and time mean of (*a*) sea surface temperature (K) and (*b*) precipitation (mm day^{-1}) illustrating change of the Pacific intertropical convergence zone in the 21st century as simulated by 19 CMIP5 RCP4.5 models. The current climate is calculated as the average during 2006-15 (contour, black for zonal mean) and the future climate as the average during 2089-98 (color, red for zonal mean). The patterns show the deviations from the tropical Pacific (30°S-30°N, 120°E-80°W) mean.