

Codimension-Two Free Boundary Problems



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Chapter 1

Introduction

The American Monsoon System (AMS) is the main source of rainfall for the most part of Latin America and the southwestern United States, regions where agricultural activity is economically crucial. Changes to the amount, timings and location of rainfall over different temporal scales has direct consequences for the livelihood of the residents in these monsoonal regions. Increasing the understanding of the mechanisms that cause short or long-term changes to the amount of rainfall in this monsoon would render key information for better predictions, which would ultimately be paramount for society in these regions.

1.1 Motivation

Temporal changes to rainfall are important for society across many scales, but the relevance of the timing and strength of rainfall is emphasized in monsoon regions. In the AMS, changes to rainfall on inter-annual scales can produce long-lived droughts that are associated with crop loss and forest fire intensification. But shorter-term changes can also produce profound effects such as flash floods and intense rainfall periods that cause crop loss via flooding. Monsoon research is thus focused on understanding the causes for change of precipitation across temporal and spatial scales.

The AMS was only recently recognized as a monsoon, as the definition for a monsoon changed from a dynamical to an agronomical definition that recognizes the seasonality of precipitation as the dominant feature of a monsoon. This fact, however, means that our understanding of the AMS is more limited compared to other monsoons given the relatively lower number of studies on the AMS when compared to other monsoons such as the Asian monsoons, that have been studied since the late 1700s. For this reason, several primary questions about general aspects of the AMS

remain open and many aspects remain unresolved. One example is the seasonal cycle of rainfall in the southern Mexico and Central America, which has defied years of research into the basic physical mechanisms that can explain the seasonal variation of rainfall. Recent attention has also increased to understand non-linear responses of precipitation of the AMS to teleconnections from the El Niño-Southern Oscillation (ENSO), which occurs just on the western coast of the continent. It is noteworthy even though the ENSO phenomena was well-known by Peruvian fishermen for centuries, the reasons for the observed different responses to the same long-distance ocean-atmosphere forcing from the Pacific Ocean remain unclear, in spite of the well documented impacts of ENSO which has caused many disastrous floods and droughts across the region.

One key tool to understand the causes and mechanisms for regional changes to monsoon are general circulation models (GCMs). These models are used to understand the role of orography, vegetation-atmosphere and ocean-atmosphere feedbacks and impacts over many aspects of Earth's climate including monsoons. However, as with other aspects of monsoon research, the use of GCMs to address key questions of the AMS has been scarce. In fact, detailed accounts of the biases –differences between a model's simulation and the real world – are rarely done with explicit emphasis on the AMS. In other words, GCMs are rarely evaluated in the AMS, so both the understanding of the biases relevant for the AMS as well as the use of GCMs to address scientific questions of the AMS is scarce.

This thesis focuses on the American Monsoon System and the outstanding questions regarding the climate variability and teleconnections affecting this monsoon. Due to both the lack of literature on the AMS in general, and specifically in the GCM studies, this thesis first evaluates a state-of-the-art climate model and using the results from this evaluation proposes several topics to investigate further in the remaining parts of the thesis.

1.2 Thesis aims and outline

This thesis uses a state-of-the-art GCM to address key outstanding questions about the variability and teleconnections of the AMS.

The key aims of this thesis are:

1. Characterize the biases of a GCM that are relevant for the AMS using a process-oriented diagnostic approach. Evaluate the roles of the role of large-

scale biases, horizontal resolution and the use of chemistry for regional monsoon representation.

2. Evaluate the seasonal variability of the monsoon onset, withdrawal and intra-seasonal changes in the model.
3. Describe and investigate the physical mechanisms associated with the seasonal cycle of rainfall.
4. To investigate the role of stratospheric-tropospheric coupling in the tropics.

The remainder of this thesis is structured as follows:

- Chapter 2 provides a review of the literature on key aspects of the American monsoons. The chapter begins by introducing the concepts of monsoons, their different physical interpretations as a global phenomena and the place of regional monsoons in the global scale. Then, the North and South American monsoons are introduced, followed up by a description of rainfall in Central and the literature on a unique monsoon-like regime of precipitation in this region. El Niño-Southern Oscillation is introduced and their impacts over North and South America are described with key emphasis on questions. Finally, the role of the stratosphere for tropical convection has recently regained attention, so this chapter ends with an up-to-date survey on the literature with specific emphasis on the stratospheric quasi-biennial oscillation.
- Chapter 3 evaluates two state-of-the-art CMIP6 models from the Met Office Hadley Centre for their representation of the AMS. The chapter describes large-scale biases over the tropics to then present local scale biases in the precipitation amount and seasonality in key regions of the AMS. The representation of ENSO teleconnections is also explored. The majority of this work has been published in *Weather and Climate Dynamics* García-Franco et al. (2020).
- During the Chapter 4 provides a method that is able to better characterise the MSD timings and strengths, as a way of analysing the mechanisms of the MSD in observations and models, analysis that is done in chapter 5. The Quasi-biennial Oscillation is proposed to be responsible for the different ENSO teleconnections shown in chapter 3 and are thus further explored using modelling experiments in chapter 6.

Chapter 2

Background

This chapter summarises the main aspects of the tropical circulation and of the global monsoon. The American Monsoon System is introduced and detail is given on the Midsummer drought of southern Mexico and Central America and El Niño Southern Oscillation teleconnections to this monsoon. Finally, a summary of the literature on the role of stratospheric-tropospheric coupling in the tropics for monsoon variability is given.

2.1 The tropical circulation and the global monsoon

Tropical climate is a result of the stronger incoming solar insolation near the equator than in poleward latitudes. This differential heating between the tropics and higher latitudes drives a meridional transport of energy by the ocean-atmosphere system. The strong incoming solar radiation warms the tropical oceans, which together with near-surface wind stresses produce large evaporative fluxes that create a very moist boundary layer and trigger deep convection. The tropical circulation can be described to a first order through the zonal and meridional circulations known as the Hadley and Walker cells that result from localized deep convective circulations.

The Hadley cell is the meridional overturning circulation that arises from the differential heating between the tropics and the midlatitudes. The Hadley cell is characterized by ascending motions in the tropics and descending motions in the subtropics, and acts to transport heat poleward from the equator (Lorenz, 1967). This Hadley cell migrates meridionally with the seasonal cycle, the winter and summer cells interact with each other but also with the midlatitudes through eddy momentum fluxes (Bordoni and Schneider, 2008). The Hadley cell is not zonally symmetric; the

boreal summer Hadley cell, for instance, is primarily a result of ascent in the Indian Ocean and the west Pacific regions with a minor contribution from ascending motions in Central and North America (Hoskins et al., 2020).

The Walker circulation is the zonal overturning circulation found in the equatorial Pacific Ocean characterized by ascending motion over the West Pacific and descending motions over the East Pacific(Walker, 1924; Bjerknes, 1969; Gill, 1980). The dynamic and thermodynamic effects of the location and strength of convection associated with the Walker circulation have strong impacts across all the tropics and also the extratropics, known as teleconnections (Cai et al., 2019).

The Inter-tropical Convergence Zone (ITCZ) is a tropical band of convective clouds and precipitation that migrates meridionally with the seasons (Schneider et al., 2014). The ITCZ is arguably one of the most relevant features of tropical climate due to the strong influence on the low- and upper-level circulation associated with ITCZ, the high tropospheric heating due to deep convection in the ITCZ and the largest precipitation rates in the tropics are found in the ITCZ. The ITCZ is characterized by a strong

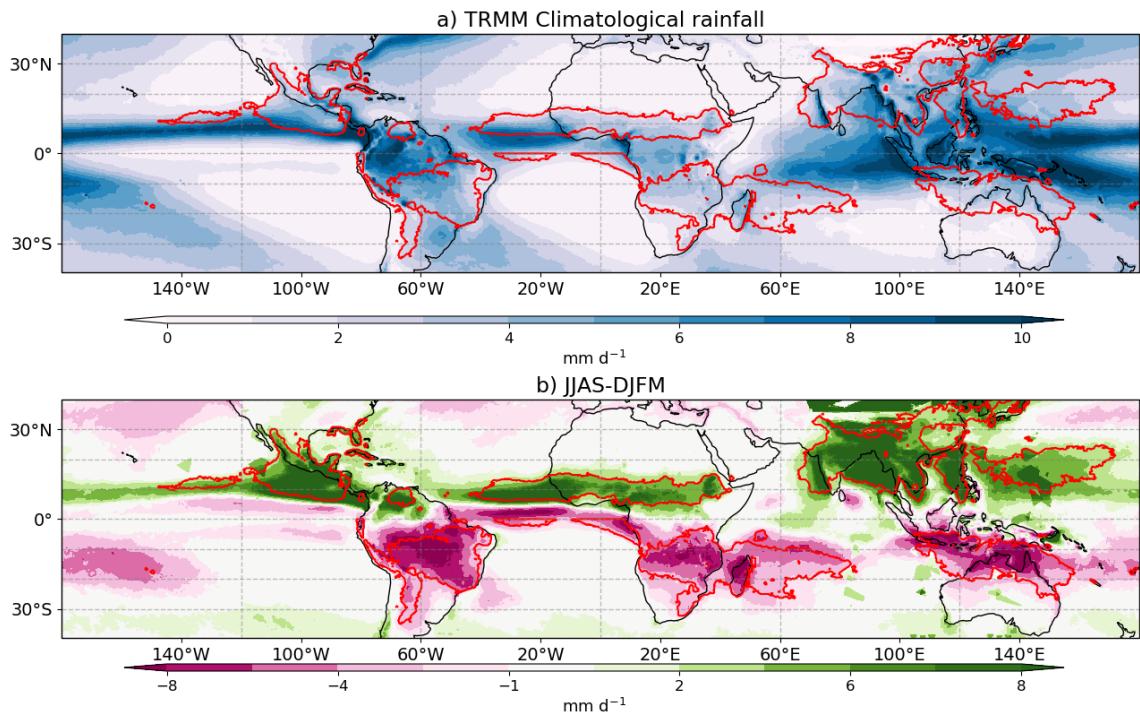


Figure 2.1: a) Climatological mean annual rainfall rates in the tropics using data from the Tropical Rainfall Measurement Mission (TRMM) dataset (1999-2018). b) The mean rainfall rate difference between boreal summer (JJAS) and austral summer (DJFM). The red contours highlight the regions where the mean summer rainfall amount accounts for more than 55% of the mean total annual rainfall accumulation.

convergent flow in the low levels and a strong divergent flow at upper levels. The meridional migration of the ITCZ, as well as the mean latitude of the ITCZ, results from the energy and momentum balances so that the ITCZ is predominantly north of the equator because of the inter-hemispheric temperature contrast (Donohoe et al., 2013; Bischoff and Schneider, 2016).

The first widely accepted view of a monsoon was that of a large-scale land-sea breeze associated with the differential warming of the land and the ocean that force a seasonal reversal of the low-level wind flow (Halley, 1687). However, this traditional land-sea breeze view of monsoons has recently been replaced by several alternative theories, an ITCZ-monsoon zonal-mean energetic framework, a convective quasi-equilibrium interpretation, the overturning Hadley cell circulation, and the moist static energy (MSE) budgets (Biasutti et al., 2018; Hill, 2019; Geen et al., 2020).

The first framework explains monsoons as a poleward extension of the ITCZ into land generalizing all monsoons as an expression of global tropical convergence resulting from the energy balance (Chao and Chen, 2001; Gadgil, 2018). This interpretation has led to the concept of *the global monsoon*, a term that encompasses all the regions in the tropics that exhibit a strong seasonality in precipitation (Zhou et al., 2016; Gadgil, 2018). In practice, the global monsoon refers to the those regions of the planet where more than 70% of the total annual rainfall falls during the summer season (Zhou et al., 2016; Wang et al., 2017).

Figure 2.1 shows the global monsoon as depicted by the TRMM dataset. By this definition, the majority of the regions over land between 5 and 10 degrees away from the equator are part of the global monsoon. A regional monsoon, such as the Indian Monsoon, is then a subset of the global monsoon with unique regional characteristics that shape this monsoon different to other regional monsoons in terms of the seasonality, the strength and the dynamics. The American Monsoon System is then the regional monsoon that is located in the subtropics of North and South America.

Bordoni and Schneider (2008) provide an alternative conceptual view of monsoons, describing the characteristic rapid onset of a monsoon as a regime transition of the Hadley cell from a edddy-momentum driven circulation, which resembles a canonical ITCZ regime, to a thermally direct circulation which resembles a monsoon-like circulation. The zonal mean MSE meridional gradient drives the ITCZ location and determines the strength of the overturning circulation by modulating the ventilation from midlatitude cooler and drier air in a feedback mechanism (Geen et al., 2020).

Event though this study used an axisymmetric framework in an idealized model, their results were broadly applicable to the Asian monsoon circulation.

Convective quasi-equilibrium (CQE) is a theory for moist convection where convection sets the vertical temperature and moisture profiles to a convectively neutral state, thereby setting the free tropospheric temperature (Neelin, 2007). For a monsoonal circulation, this theory emphasizes the coupling of convection and dynamics predicting that the subcloud layer equivalent potential temperature maxima must be collocated with the free tropospheric saturation equivalent potential temperature (Nie et al., 2010; Geen et al., 2020). The rapid onset of the Asian monsoon has been shown to be associated with the boundary layer moist entropy distribution, in agreement with predictions of CQE (Nie et al., 2010; Boos, 2015; Ma et al., 2019).

Several studies examine the monsoon phenomena through an axi-symmetric framework that assummes zonal symmetry and aims to understand the large-scale dynamical influence over the extent and strength of the axi-symmetric monsoon through global energetic diagnostics (e.g. Faulk et al., 2017; Geen et al., 2019; Byrne and Zanna, 2020). The zonal-mean framework is common to the Hadley cell interpretation of monsoons (Bordoni and Schneider, 2008), as well as the ITCZ-monsoon theory. However, regional monsoons are shaped by the asymmetries imposed by the orography, the characteristics of the surrounding ocean basins, land-sea contrasts and also the role of vegetation-hydrology coupling (Wang et al., 2017; Pascale et al., 2019). The importance of zonal asymmetries has raised multiple issues with large-scale so-called monsoon dynamics theories, as several predictions of these theories are not consistent with observations of regional monsoons (e.g. Nie et al., 2010; Smyth et al., 2018; Biasutti et al., 2018; Pascale et al., 2019).

Recent reviews acknowledge that all these frameworks have significant shortcomings to be applied to regional local monsoons (Biasutti et al., 2018; Hill, 2019; Geen et al., 2020). These reviews conclude that a framework that reconciles the global energetic perspective with the characteristics of regional monsoons would be crucially important and very useful, but as several authors point out (e.g. Biasutti et al., 2018; Hill, 2019), also very hard to formulate. For example, the North and South American Monsoons depart from CQE, as precipitation does not follow the maxima in subcloud equivalent potential temperature (Nie et al., 2010; Geen et al., 2020). One possible reason for this is that the free-troposphere over southwestern North America is significantly drier than in other monsoon regions, decoupling the free troposphere from the boundary layer. One alternative hypothesis is that ventilation of low moist

entropy air from the midlatitudes is responsible for this decoupling in the American monsoons (Boos, 2015).

The MSE budget framework suffers both from theoretical and practical shortcomings. One practical shortcoming is that the calculation of the budget terms post hoc in reanalysis or models results in very large residuals (Hill, 2019), so these frameworks work best when the calculations are done inside the budget terms to be integrated online at each time-step (e.g. Ma et al., 2019). The theoretical shortcoming is that the surface fluxes over land, e.g., in the Sonoran and Saharan deserts and the deep Amazon make the estimations of the roles of hydrology-vegetation feedbacks and their potential contributions to the MSE budget in observations very difficult to assess (Boos and Korty, 2016; Pascale et al., 2019). The use of simpler moisture budgets has proven useful in a regional monsoon context to investigate the sources of moisture for a monsoon in current (Ordoñez et al., 2019; Martinez et al., 2019) and future climates (Smyth and Ming, 2020), but this budget is mostly a tool and not a coherent theory for process-level understanding of monsoons.

The Hadley cell interpretation of monsoons has significant shortcomings to depict some regional monsoons, particularly those that are not the Asian monsoon as the overturning circulation in the South Asian monsoon is strong enough to be represented by a clear thermally direct regime. However, this energetic framework assumes no zonal transport of energy, which minimizes the role of orography and land-sea interaction (Biasutti et al., 2018). . One might reasonably infer from these results that the timing of transition in zonal mean overturning cells would be similar for monsoons at different longitude but similar latitudes, which is not the case (Wang et al., 2017). Furthermore, a monsoon restricted to a small area, such as a the North American and African monsoons may not present a clear zonally averaged overturning regime, and may be significantly affected by local zonal circulations (Zhai and Boos, 2015). For instance, Smyth et al. (2018) shows that the simulated West African monsoon when forced with different solar forcings exhibits a decoupling between the zonal-mean ITCZ location, the strength of the local Hadley cell and the monsoon rainfall, in clear contrast to the predictions of this framework (Bordoni and Schneider, 2008).

In short, despite significant progress in our understanding of the monsoon phenomena at the planetary-scale through zonal mean energetic frameworks, there is an important gap between large-scale theories of monsoon dynamics and the observed regional monsoons. The next section presents a summary of the American Monsoon literature, which explains the characteristics of these monsoons through the effect of

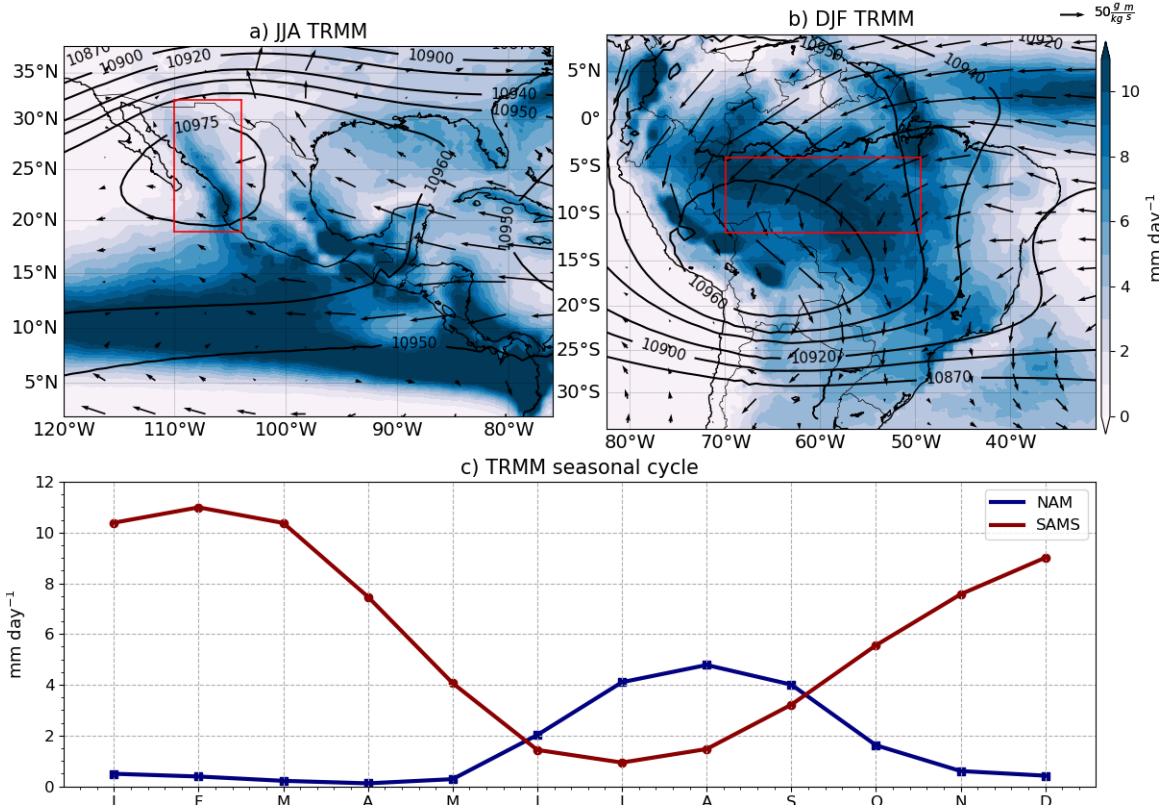


Figure 2.2: Climatological mean a) boreal and b) austral summer rainfall (shading), 850 hPa moisture flux (vectors) and geopotential height at 250 hPa (contours) in a) southern North America and b) South America. c) Monthly-mean seasonal march of precipitation in the TRMM dataset for two area-averaged time-series, the North American Monsoon (NAM) and the South American Monsoon System (SAMS) shown in the red rectangles in a-b).

regional features and dynamics, seemingly detached from the literature in this section. The AMS literature is therefore, seemingly, detached from the literature in this section.

2.2 The American Monsoon System

The American Monsoon System (AMS) is the main source of rainfall for tropical Latin America and is typically subdivided into the North and South American monsoon systems (Vera et al., 2006). The North American Monsoon is found in south-western North America, see Figure 2.2a, extending north from central-west Mexico into the southwestern United States, (Adams and Comrie, 1997; Stensrud et al., 1997; Vera et al., 2006). The seasonal cycle of rainfall in the North American Monsoon is characterised by a wet July-August-September season and significantly drier conditions

during the rest of the year (Adams and Comrie, 1997) (Figure 2.2c).

Three temporal stages describe the evolution of the North American Monsoon (Adams and Comrie, 1997; Geil et al., 2013). First, the onset stage (May-June) starts with a strong surface warming that leads to very high temperatures in the desert region. Simultaneously, the sub-tropical jet weakens and migrates north decreasing the frequency of mid-latitude disturbances in the monsoon region (Douglas et al., 1993; Turrent and Cavazos, 2009). These factors combine to develop a low-level (upper-level) thermal surface low (anticyclone) and moisture influx from the nearby Gulf of California and easternmost Pacific Ocean (Douglas et al., 1993; Geil et al., 2013). Maturity (July-August) is the peak period of monsoon rainfall characterised by sustained deep convection (Barlow et al., 1998) and significant increases in low and mid-level moisture flux convergence and mid-level latent heating (Adams and Comrie, 1997; Cook and Seager, 2013). This latent heating caused by deep convection can be diagnosed in the upper-level geopotential height (Figure 2.2a) in the form of an anticyclone centred on the monsoon region.

The moisture flux convergence decreases in August, after which precipitation recycling (Dominguez et al., 2008) plays an important role in keeping deep convection active until September. Decay (September-October) is the last stage of the monsoon, in many ways opposite to the onset stage, as is characterised by the equatorward migration of the sub-tropical jet (Higgins et al., 1997; Geil et al., 2013), evaporation in the nearby basins decreases and deep convection in the monsoon region gradually disappears (Douglas et al., 1993).

The origin of the high levels of moisture at low and midlevels in the monsoon region has been a matter of debate for a long time (Adams and Comrie, 1997; Barlow et al., 1998; Vera et al., 2006; Ordoñez et al., 2019). A large number of studies acknowledge that the main source of moisture for the North American Monsoon is the East Pacific Ocean and to a second order, mid-level moisture advected from the Gulf of California can mix in the column (e.g. Adams and Comrie, 1997; Stensrud et al., 1997; Vera et al., 2006; Turrent and Cavazos, 2009; Ordoñez et al., 2019).

The South American Monsoon is a primary source of precipitation for South America, especially in the Amazon region (Gan et al., 2004; Vera et al., 2006; Jones and Carvalho, 2013). During austral summer (DJF), monsoon rainfall accounts for over 60% of the total annual precipitation in the Amazon (Gan et al., 2004; Marengo et al., 2012), whereas austral winter rainfall accounts for less than 5% of the total annual rainfall (Vera et al., 2006). In the central Amazon, convective precipitation is observed from early October but the main rainy season extends from December to

April (Machado et al., 2004; Adams et al., 2013), whereas convection in southeastern Brazil and Paraguay starts in November and peaks in January and February (Marengo et al., 2001; Nieto-Ferreira and Rickenbach, 2011).

A surface heat low appears in Bolivia in early austral summer, known as El Chaco Low, as a result of strong warming in austral spring (Marengo et al., 2012; Sulca et al., 2018). As this surface heat-low strengthens, low-level convergence drives the circulation into the low region. Simultaneously, an upper-level anti-cyclone (Fig. 2.2b), known as the Bolivian High, develops in the same region as a signature of strong deep convection and latent heating (Marengo et al., 2001; Vera et al., 2006).

This low-level wind circulation importing moisture from the Atlantic is one of the most important features of the SAMS (Marengo et al., 2012; Wang et al., 2017) as the flow modulates the moisture flux to the mainland and influences the occurrence of active and break phases of the SAMS (Jones and Carvalho, 2002), as well as changes in the temporal and spatial distribution of rainfall (e.g. Giannini et al., 2004; Bombardi and Carvalho, 2011).

2.3 A review on the proposed mechanisms for the Midsummer drought

The characteristics of the seasonal cycle of precipitation in northwestern Central America and southern Mexico fit the definition of a monsoon climate (Wang et al., 2017). characterized by the majority of precipitation observed during local summer. However, this region shows a unique climatological precipitation feature. After monsoon onset, rainfall decreases considerably around the midsummer; this decrease is followed by a secondary increase in precipitation in the late summer (Mosiño and García, 1966), and for this reason this feature of the seasonal cycle is most commonly referred to in the literature as Midsummer drought (MSD) (Magaña et al., 1999).

The intraseasonal variations of precipitation associated with the MSD have been known for centuries, as ancient mayan texts suggest that agricultural rituals that asked for rain-bearing clouds to the gods had a similar seasonality to that of the MSD (Jobbová et al., 2018). In current days, the MSD is well known by local farmers who refer to the drier midsummer period as ‘El Veranillo’ in Central America and ‘canícula’ in southern Mexico because the drier period coincides with the Canis Major constellation appearing in the sky (Dilley, 1996), so that the MSD variations have shaped agricultural practices in the region for centuries.

The two peak structure of the MSD has been diagnosed in the observed climatological precipitation of several regions of Mexico, El Salvador, Belize, Guatemala, Costa Rica and Cuba (e.g. Mosiño and García, 1966; Magaña et al., 1999; Durán-Quesada et al., 2017; Perdigón-Morales et al., 2018). However, notable differences in the seasonal cycle of precipitation have been found between the mainland Central America and the Caribbean islands. The so-called first peak of precipitation occurs in May in Cuba and in June in northern Central America whereas the second peaks are observed in October and September for the mainland and the Caribbean, respectively.

In spite of extensive research to understand the physical mechanisms associated with the MSD (e.g. Magaña et al., 1999; Giannini et al., 2000; Gamble et al., 2008; Herrera et al., 2015; Maldonado et al., 2017; Straffon et al., 2019), debate remains over which is the leading-order mechanism that causes rainfall to decrease at midsummer and increase again at the end of the summer. Fundamental questions remain unclear such as whether the MSD is caused by two precipitation enhancing mechanisms (Karnauskas et al., 2013) or a mechanism that inhibits rainfall at midsummer (Durán-Quesada et al., 2017). Furthermore, the association between the MSD in Central America and in the Caribbean is still disputed (Gamble et al., 2008), as most studies suggest that the two regimes are unrelated and therefore two different explanations are required to account for the two MSDs in these regions.

Any complete theory or conceptual model must account for the following characteristics of the seasonal cycle. First, the theory must explain the timing and strength of the first peak of rainfall. Second, the timing and strength of the MSD, i.e., what causes rainfall to decrease at midsummer. Finally, the theory must explain the timing and mechanism driving the second increase in precipitation after the midsummer.

One of the first theories to account for the bi-modal distribution of the seasonal cycle of rainfall was the double crossing of the ITCZ (Hastenrath, 1967). However, this theory fails to explain the MSD signal seen at latitudes as high as 25°N (Perdigón-Morales et al., 2018; Zhao et al., 2020) which is further north than the northernmost extension of the ITCZ (Schneider et al., 2014), and certainly the ITCZ does not cross twice so far from the equator.

Magaña et al. (1999) and Magaña and Caetano (2005) proposed a mechanism driven by radiative-convective feedbacks between the East Pacific sea-surface temperatures (SSTs) and deep tropical convective clouds. The height and strength of convection, the incoming shortwave and the SSTs are strongly coupled in their framework. The EP SSTs peak in May triggering large evaporative fluxes and deep convection in the EP ITCZ and Central America. The high convective clouds produce a

radiative cooling effect at the surface due to decreased incoming shortwave radiation. This cooling decreases SSTs and deep convective activity and thus accounts for the modest decrease in rainfall during the midsummer. The second peak in September is driven by the the decreased frequency of tall convective clouds during the MSD period in July and August, which act to reduce the cooling effect of the anvil clouds and increases incoming shortwave, SSTs and surface fluxes, and eventually increasing precipitation for late August and September (Magaña et al., 1999).

Other studies suggest the seasonal evolution of North Atlantic Subtropical High (NASH) and the associated geostrophic flow are the primary cause of the bi-modal regime, particularly for the MSD in the Caribbean (e.g. Mapes et al., 2005; Gamble et al., 2008; Curtis and Gamble, 2008). The NASH is a subtropical anticyclone in the Atlantic Ocean that shifts southwest early in boreal summer. The expansion and intensification of the NASH in boreal summer, according to this theory, strengthens the low-level trade winds, controlling the seasonal cycle of a low-level jet known as the Caribbean Low-Level Jet (CLLJ).

The CLLJ is key for regional variability and climate of the Caribbean, northern central America and southern United States, because the strength, height and direction of the flow in the Caribbean plays a key role for the regional moisture transport (Giannini et al., 2000; Martinez et al., 2019; García-Martínez and Bollasina, 2020). The expansion of the western flank of the NASH is argued to strengthen the CLLJ which cools the SSTs, through the effect of wind stress and mixed-layer mixing (Gamble et al., 2008; Martinez et al., 2019). The cooling of SSTs diminishes evaporation and therefore low-level moisture which leads to less precipitation, at least locally. Because the expansion of the NASH is closely aligned with the timing of the MSD in the Caribbean, the NASH effect on the CLLJ is argued to generate the bimodal regime of precipitation (Gamble et al., 2008; Martinez et al., 2019).

The easterlies crossing from the Caribbean Sea to the East Pacific (EP) and their associated moisture flow are recognized as an important driver as the strength of the easterlies modulates ascending and descending motions through vertical wind shear and moisture divergence (Herrera et al., 2015; Corrales-Suastegui et al., 2020; Zhao et al., 2020). In particular, moisture budgets have been used to diagnose the changes to the regional and temporal distribution of moisture (Martinez et al., 2019). The intra-seasonal variability of precipitation have thus been explained in terms of the variability of the easterly wind flow in the CLLJ and the EP Ocean (Herrera et al., 2015; Martinez et al., 2019; Zhao et al., 2020), with the hypothesis that a stronger CLLJ will force a stronger MSD signal, i.e., suppress precipitation in the midsummer.

The SST gradient between the EP Ocean and the Caribbean Sea has been argued to modulate the moisture flow across the continent on interannual time-scales (Martinez et al., 2020). This is particularly relevant as the Pacific Ocean is projected to warm more than the Caribbean Sea in future decades, which will change the SST gradient, strengthen the CLLJ and shift the regional precipitation patterns (Corrales-Suastegui et al., 2020).

Herrera et al. (2015) shows that during the drier months in Central America, stronger convective activity is found west of the Central American coast. This evidence suggests that the coupling of EP SSTs to the gap flow that originated from the CLLJ in the Caribbean Sea controls the location of ascending and descending motions, thereby explaining some features of the Central American MSD. Herrera et al. (2015) further argued that the exit region of the CLLJ is located to the east of the region of strongest MSD signal, which suggests that the moisture divergence effect over the central American MSD is minimal.

A different mechanism, proposed by Karnauskas et al. (2013), argues that the biannual crossing of the solar declination angle can control precipitation to the extent of explaining the bimodal characteristics of the seasonal cycle. In this mechanism, the MSD is driven by two precipitation enhancing periods that are separated by a relatively normal, and drier, period. This theory differs from those previously discussed which explained the MSD through mechanisms that inhibit convective activity in the midsummer whereas Karnauskas et al. (2013) argues that the solar declination angle that crosses twice through Central America, once during June and a second time during September, increases convective activity during each crossing.

The variations of incoming shortwave radiation associated with the declination angle modulate the SSTs, surface fluxes and therefore convective activity. In other words, the first crossing of the solar declination angle increases the incoming shortwave radiation which increases the SSTs, evaporation and precipitation, i.e., the first peak. The second crossing, similarly, explains the second peak as the second increase in incoming shortwave promotes more deep convection than during the MSD.

Other mechanisms have been proposed arguing that the MSD is a result of the result of vertical wind shear affecting convective instability or the Saharan dust controlling the microphysics of clouds (Angeles et al., 2010). For instance, Perdigón-Morales et al. (2019) also finds a link between the frequency and spatial distribution of the first peak rainfall rates and the Madden-Julian Oscillation.

2.4 El Nio Southern Oscillation: impacts to the American monsoon system

El Nio-Southern Oscillation (ENSO) is a phenomena that primarily affects the local ocean and atmosphere of the equatorial Pacific Ocean, but these changes are profoundly important for the global climate system, which is why ENSO is commonly known as the leading mode of interannual variability. The term '*El Nio*' was initially coined by Spanish colonizers when they learnt from Peruvian fishermen that the ocean surface temperatures in the easternmonst Pacific Ocean increased notably in some years around December time. For religious reasons, the colonizers termed the SST increase as Christ Child – *El Nio*.

Later on, sir Gilbert Walker (1924) coined the term *Southern Oscillation* to describe the synchronous changes to the sea-level pressure of the Indo-Pacific region and South America. Walker (1924) and Walker (1932) are the first analyses of synchronous effects of the tropical circulation over local precipitation, temperature and pressure. Further research (e.g. Troup, 1965) would highlight that these remote changes in pressure were driven by the east-west pressure gradient in the equatorial Pacific.

The changes in the pressure field associated with the Southern Oscillation (SO) are now part of what is known as the Walker circulation, which intertwines the dynamics of the zonal circulation in the East Pacific with the SSTs over the underlying ocean. ENSO is then characterized as a coupled phenomena composed of an oceanic part, *El Nio*, and an atmospheric component associated with the zonal circulation but best characterized by changes to the surface pressure field, the Southern Oscillation.

ENSO has several unique features, such as no robust periodicity as events may occur every 2 to 7 years and a seasonal phase-locking that are associated with ENSO events peaking in boreal winter in observations (Wang and Picaut, 2004). Even though the underlying physics that cause ENSO and explain the variability in the periodicity of the phenomena is still debated (Wang and Picaut, 2004; Christensen et al., 2017), several aspects are now better understood. For example, the local effect that ENSO events have over on the location and strength of deep convection in the equatorial Pacific have long been thoroughly described (Trenberth, 1997; Neelin et al., 1998).

During a neutral state of ENSO, the Walker circulation is found in the climatological state, with ascent and wet conditions in the West Pacific and descent and drier conditions in the East Pacific. During El Nio the Walker circulation and low-level trade winds weaken which is associated with an eastward shift of deep convection

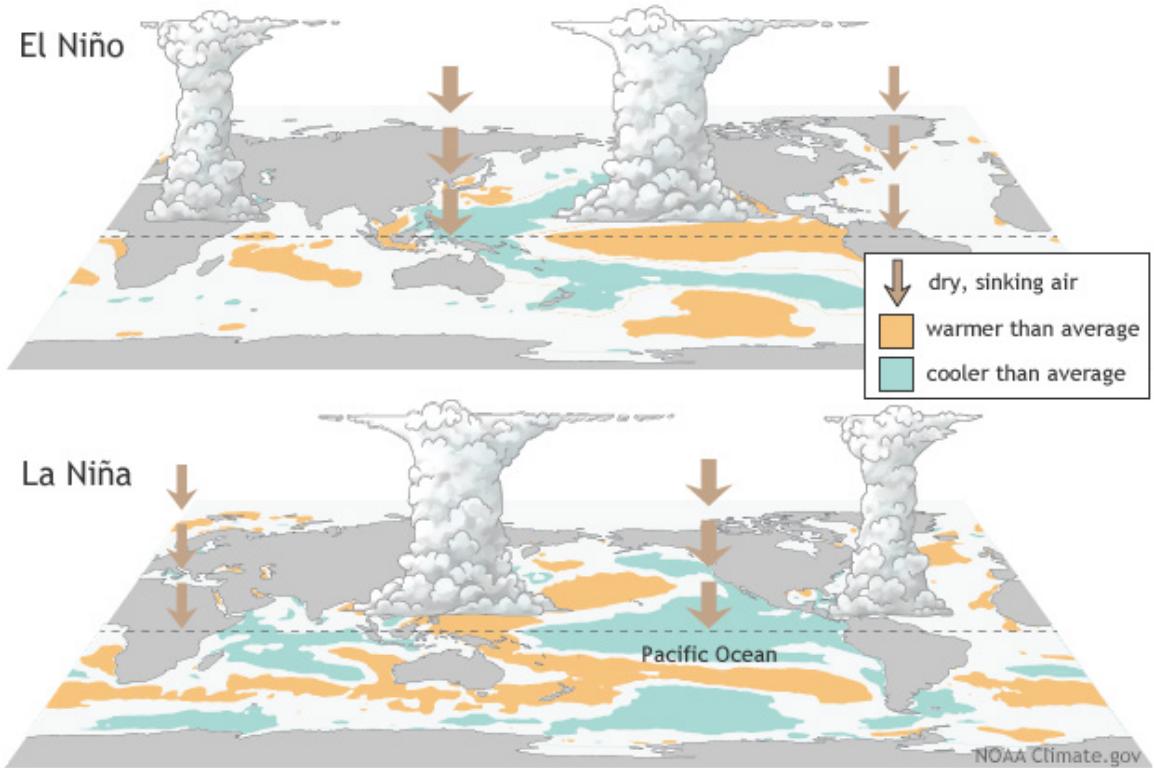


Figure 2.3: Schematic of the positive (upper) and negative (lower) phases of ENSO. Regions with tall clouds indicate more ascent and convection than normal whereas brown arrows indicate dry descending air. Obtained from the National Oceanic and Atmospheric Administration at <https://www.climate.gov/enso>.

along the equatorial Pacific (Figure 2.3), with convective rainfall becoming more frequent in the central and even eastern Pacific than normal (Neelin et al., 1998; Wang and Picaut, 2004). During La Niña the opposite happens and the Walker circulation strengthens which leads to stronger convection in the West Pacific and stronger ascent on the East Pacific (Figure 2.3).

In other words, ENSO imposes a strong control on the location and strength of the Walker circulation (Figure 2.3). These changes to the strength and position of the convective regions in the Pacific Ocean can then propagate to other regions of the planet; these far-distant effects are commonly known as *teleconnections*. For example, ENSO has a direct effect over other tropical regions outside of the Pacific through the Walker circulation, see Figure 2.3, as upper-level wind anomalies induce anomalous vertical motions over the monsoons in West Africa (Ropelewski and Halpert, 1986, 1987) or South America (Sulca et al., 2018). Other mechanisms of ENSO teleconnections to higher latitudes include changes to the position and strength of sub-tropical jets (Fereday et al., 2020), changes to the Pacific and North American circulation pat-

terns (Bayr et al., 2019) as well as impacts to the North Atlantic via the stratospheric polar vortex (Domeisen et al., 2019).

In South America, the effects of ENSO are felt throughout the continent and throughout economic sectors from Peruvian fishermen (Takahashi, 2004) to the Amazon rain-forest and the plainlands in South-eastern South America(Grimm, 2011; Marengo et al., 2012). One key aspect of current research on ENSO impacts to South America is the observed non-linearity and non-symmetry in the teleconnections, which has mainly been attributed to ENSO diversity (Tedeschi et al., 2015; Cai et al., 2020). A non-linear teleconnection refers to a non-linear scaling between the strength of an ENSO event, typically measured by an SST index, and the magnitude of the response, in most cases precipitation response.

Observations have shown that the maximum SST anomaly does not always appear in the same region of the Pacific Ocean (Ashok and Yamagata, 2009; Dommenget et al., 2013). These differences in the SST patterns are referred to as ENSO *diversity* or *flavours* which can be broadly summarized as two flavours for each phase, based on the location of the SST anomaly these events are commonly separated into Central and Eastern Pacific events. In observations, each type of event is usually also associated with the strength of the event (Dommenget et al., 2013).

The strength and patterns of ENSO teleconnections have been shown to depend on the type of event for the South American monsoon (Rodrigues et al., 2011; Sulca et al., 2018). Cai et al. (2020) provides a recent review on the differences in the impacts that Central and Eastern Pacific (CP and EP) events have on South America. The observed record shows that the teleconnections affecting the Amazon and northeastern Brazil are most pronounced during EP El Nio events and CP La Nia events than the CP El Nio events and EP La Nia events. This recent review also highlights the need for further modelling work to test observation-driven hypothesis, as the observed record is too short to make confident statements about the mechanisms associated with ENSO teleconnections.

2.5 Stratosphere-Troposphere Coupling in the Tropics

The troposphere is the lowermost layer of the atmosphere ranging from the surface up to 10-15 km where the vertical temperature profile is characterized by a decrease of temperature with height. The layer above the troposphere is called the *stratosphere* where the vertical temperature gradient is reversed and temperatures increase with height,

this layer and spans from 10-20 km up to 50 km in altitude (Andrews et al., 1987). The layer in between the troposphere and the stratosphere is the tropopause, which is a transition region where the vertical temperature gradient reverses and the atmosphere is stable to upward motions.

Stratosphere-troposphere coupling refers to events or processes where the two layers are notably affected by each other so that the temperature or wind flow of the two layers vary together, or *couple*. One prominent example of this coupling is the effect of the stratospheric polar vortices over the zonal flow in the troposphere (Thompson et al., 2005; Domeisen et al., 2019). However, semi-periodic climatic variability in either layer can also have effects over the other layer. The dominant mode of interannual variability in the equatorial stratosphere is such an example and is introduced in the following section.

2.5.1 The Quasi-biennial oscillation (QBO)

The stratospheric quasi-biennial oscillation (QBO) was discovered 60 years ago through balloon observations that revealed that the zonal winds reverse direction in a semi-periodic way with accompanying temperature variations (Ebdon, 1960; Reed, 1964). The QBO has then been characterized by further observations as alternating easterly and westerly wind regimes associated with a descending zonal wind shear from up to 50 km down to 10 km, with a mean oscillatory period of 28 months (Baldwin et al., 2001). The downward propagation of the easterly and westerly wind regimes, amplitude and the mean period have been explained by the interaction of a broad spectrum of gravity and Kelvin waves of tropospheric origin with the equatorial stratospheric zonal mean flow (Baldwin et al., 2001).

The wind variation in the middle stratosphere associated with the QBO are greater than the seasonal cycle (Andrews et al., 1987) and this vertical wind shear imposes a temperature signal through the thermal wind relationship, which can be expressed as:

$$\frac{\partial u}{\partial z} = \frac{-R}{H\beta} \frac{\partial^2 T}{\partial y^2}, \quad (2.1)$$

where $\partial u / \partial z$ is the vertical shear of the zonal wind, R is the ideal gas constant, y is the latitude, H is a scale height of the atmosphere (7-8 km) and β is the first derivative of the Coriolis term in the meridional coordinate y .

Through this thermal wind relationship a westerly shear would have warm anomalies beneath whereas an easterly shear would impose cold temperature anomalies. In

turn, these temperature anomalies couple to the mean meridional circulation (Baldwin et al., 2001) causing an anomaly on this circulation, an effect which is often referred as the secondary circulation of the QBO (Plumb and Bell, 1982; Li et al., 1995; Ribera et al., 2004). This anomalous circulation is characterized by reduced upwelling during westerly shear phases and increased upwelling during the easterly phase. These meridional circulation perturbations adiabatically warm (anomalous descent at the equator) and cool (anomalous ascent) for westerly and easterly shears, respectively, at the equator.

The combination of dynamic and thermodynamic effects of the QBO in the equatorial stratosphere are associated with long-distance impacts across the stratosphere (Holton and Tan, 1980; Lu et al., 2020) and down to the surface (Garfinkel and Hartmann, 2010; Gray et al., 2018). Even though the domain of the QBO is in the tropics, the better recognized effects of the QBO are in the extra tropics, particular how the direction of the zonal mean flow in the equatorial stratosphere modulates the propagation of extratropical waves and therefore also influences the wintertime stratospheric polar vortex (Lu et al., 2020), which can have profound impacts over the surface in polar and midlatitude regions.

The meridional circulation of the QBO and the associated temperature anomaly impact the height and temperature of the tropopause in the tropics (Baldwin et al., 2001; Tegtmeier et al., 2020a,b). The easterly phase of the QBO (QBOE) is associated with a higher and colder tropopause in the tropics whereas the westerly phase (QBOW) is observed with lower and warmer tropical tropopause (Tegtmeier et al., 2020a). These variations in the tropopause characteristics, amongst other effects associated with the QBO, have been hypothesized to affect, to different extents, a range of tropical phenomena, and in particular, deep convective systems.

2.5.2 Tropical teleconnections of the QBO

The influence of the QBO on the dynamic and thermodynamic characteristics of the tropical upper-troposphere-lower stratosphere (UTLS) region has raised interest in possible indirect effects of the QBO over tropical deep convection and clouds. Gray (1984) was amongst the first to suggest an influence of the QBO over tropical systems, in particular, that Atlantic tropical cyclone activity was enhanced during QBOW compared to QBOE. Gray et al. (1992) further argued that the anomalous vertical wind shear associated with the QBO affects the strength of convection in monsoonal and convergence zones to the extent that the vertical wind shear can modify ENSO

frequency. Their results suggest that El Niño events are favoured during QBOE and La Niña events are more frequent during QBOW.

The evidence by Gray et al. (1992) has motivated further observational and modelling research on QBO tropical teleconnections; some of this research has contested Gray's results (e.g. Chan, 1995; Camargo and Sobel, 2010; Hansen et al., 2016). For example, Giorgetta et al. (1999) was amongst the first to use a global climate model (ECHAM4) to investigate the effects of the QBO over tropical convection. Giorgetta et al. (1999) focused on the role that the QBO plays in modulating the strength of the East Asian and Indian monsoons. Their findings suggest that monsoon variability was partially modulated by the QBO, with the strongest effects associated with the QBO found for cloudiness at 100 hPa. Giorgetta et al. (1999) argued that these differences could be explained by the effect of the QBO on the UTLS static stability and a consequent effect over the vertical extent of deep tropical convection.

Further modelling work has been carried out, for instance by Garfinkel and Hartmann (2010) and Garfinkel and Hartmann (2011) that used the Whole Atmosphere Community Climate Model (WACCM) to understand the effect of the QBO over tropical precipitation, the subtropical jets and the wintertime polar vortex. Garfinkel and Hartmann (2010) shows that the canonical ENSO teleconnections to the North Pacific are stronger during QBO W in WACCM and reanalysis suggesting that the QBO modulates the wave propagation activity associated with ENSO events.

Garfinkel and Hartmann (2011) uses perpetual winter conditions on the WACCM model and further shows that the effect of the QBO on the upper-tropospheric zonal wind in the equator and the strength and location of subtropical jets, particularly at the exit region of the jets. Their experiments show that the QBOE-QBOW zonal wind response in the upper troposphere is zonally asymmetric at the equator but zonally symmetric in the subtropics as the subtropical jet is weakened during QBOE conditions.

The response of convection to the QBO UTLS temperature anomalies was investigated in cloud-resolving model simulations by Nie and Sobel (2015). Their experimental design used the System for Atmospheric Modelling (SAM), varying SST boundary conditions with increments over a baseline SST level of 301 K, the use of the weak-temperature gradient approximation and an idealized vertical temperature profile to simulate the effect of the QBO temperature signal. Figure 2.4 shows that the precipitation and convective features differences between QBO phases depends on the SST forcing.

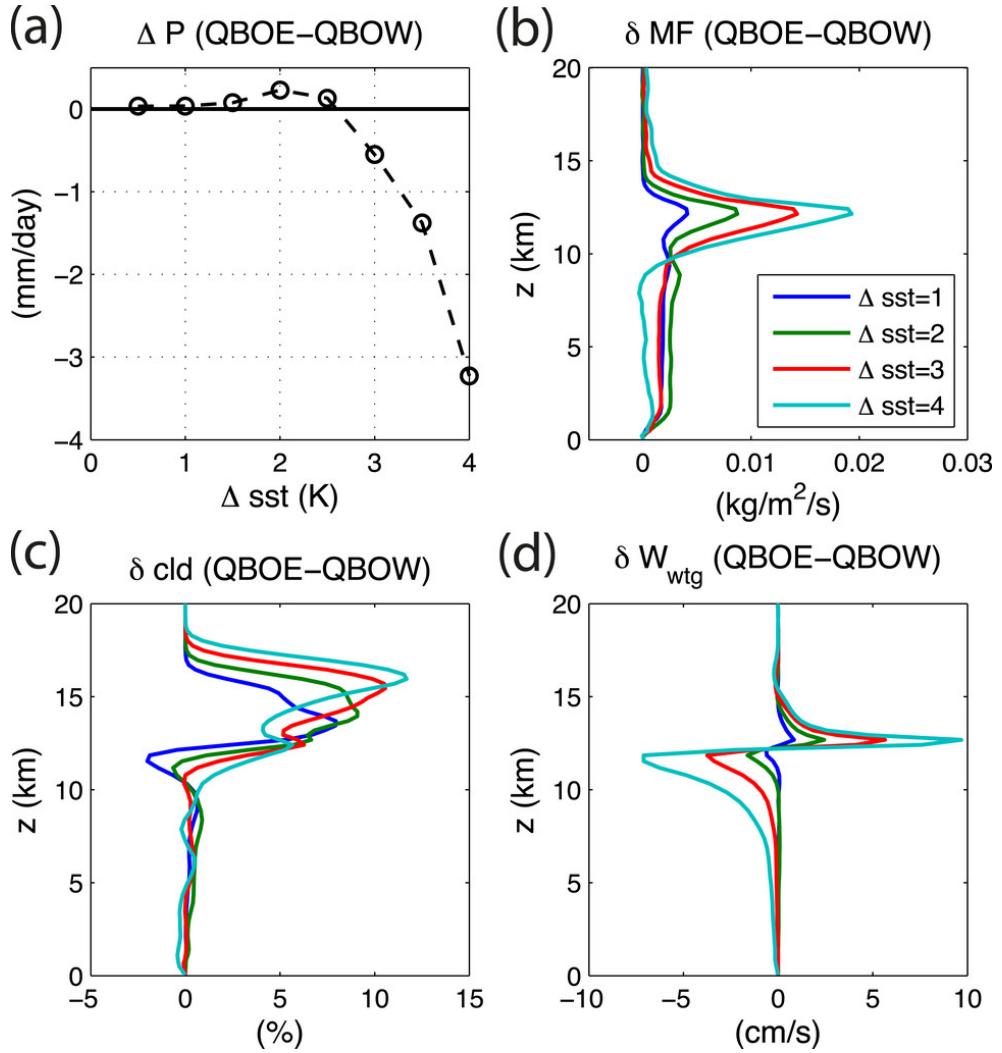


Figure 2.4: (a) QBO anomalous precipitation as a function of SST forcing. The Δsst are increments over a baseline of 301 K throughout the whole model domain. (b)(d) The differences of mass flux in cloud cores, cloud fraction, and the parametrized large-scale vertical velocity derived from the weak temperature gradient approximation, respectively, between experiments with the QBOE and QBOW temperature profiles. Figure 3 from Nie and Sobel (2015).

The precipitation difference QBOE–QBOW is positive under relatively small SST anomalies but in experiments with large SST anomalies this difference becomes negative and overall larger than under small SSTs. The difference in mass flux is also sensitive to the underlying SSTs, as increased mass flux during QBO E is increased for larger SSTs. In other words, the QBO influence on precipitation is non monotonic and largely depends on the underlying SST field.

The cloud fraction in the upper troposphere is increased during QBO E compared

to QBO W whereas the ascending motions show peculiar behaviour with stronger ascent above 12 km for QBOE and weaker ascent below 12 km. Their results suggest that the QBO influences convection in two ways that are non-linear and are competing mechanisms. In agreement with previous arguments, Nie and Sobel (2015) finds that the mas flux in the upper troposphere is increased during QBO E but also that this increases the gross moist static stability (GMS) increasing the efficiency of large-scale vertical motions during QBO E for large SSTs, which acts to reduce precipitation. Secondly, the QBO modifies the fraction of high-level clouds resulting from deep convection which modifies the radiative heating to the column which increases precipitation during QBOE. Figure 2.4a then shows how these competing effects change for different SSTs with the GMS effect dominating for large SSTs.

Various attempts have been made to determine a relationship between the QBO and deep convection in observations. One influential study by Collimore et al. (2003) analysed satellite-derived out-going longwave radiation (OLR) in the tropics composited by QBO phase. These composites suggest that OLR is significantly different between QBO phases in most monsoon regions, such as Central America and the West Pacific, with an overall indication that convective activity is reduced during QBOW compared to QBOE. This influence, however, was not found to be zonally symmetric and in fact the longitudinal variations of the QBO-related OLR differences were suggestive enough that Collimore et al. (2003) argued for a possible role for the QBO to modulate the Walker circulation, which would explain the lack of zonal symmetry in their results.

Another relevant study by Liess and Geller (2012) found that satellite-derived cloud thickness and frequency and upper-level velocity potential had a significant and longitudinally asymmetric response to the QBO. In particular, their results show increased convective activity during QBOE in the West Pacific but the opposite for the East Pacific. For this reason, Liess and Geller (2012) also argued that the strength of the tropical overturning circulation may be modulated by the QBO, indicating the possible role of both the vertical wind shear and the upper-level static stability to modulate deep convection.

The topic of QBO tropical teleconnections has regained attention due to recent findings suggesting a link between the QBO and the Madden-Julian Oscillation (MJO) (Son et al., 2017) which motivated extensive research (see e.g. Lee and Klingaman, 2018; Wang et al., 2019; Martin et al., 2020) due to the worldwide impact of the MJO. The MJO in observations shows a stronger amplitude and more predictability

during QBO E, but further inspection in cloud-permitting and forecast models have not provided conclusive answers to this puzzle (Martin et al., 2019, 2020).

Questions still arise as to whether this tropical link is real or due to chance, for instance Wang et al. (2019) argued that the increased predictability of the MJO under the QBO E phase is included in the initial conditions, and thus not a result of a mechanistic effect of the QBO on the MJO. More generally, whether the QBO has a considerable effect on deep convection in general is debated as several plausible mechanisms exist in the literature (see e.g. Nie and Sobel, 2015) such as the effect of wind shear, the tropopause height, the cold-point temperature, static stability and/or feedbacks with very high cirrus and cumulonimbus clouds.

Chapter 3

Data and methods

3.1 Observations and reanalysis data

Continuous and reliable observations of the atmosphere and ocean have only been possible until recent decades, most of which rely on the use of satellite data one way or another. This thesis uses several data sources for precipitation from various gridded precipitation datasets. The rest of dynamic diagnostics used in this thesis are from the newest reanalysis from the European Centre from European Centre for Medium-Range Weather Forecasts, described below. Table 3.1 summarises relevant information of the observations and reanalysis datasets used in this study.

3.1.1 Gridded precipitation datasets

The Tropical Rainfall Measurement Mission (TRMM) dataset is a multi-satellite multi-sensor infra-red precipitation product that is available on several versions that are made with different algorithms and calibrations with surface rain-gauge data (Huffman et al., 2007). This thesis uses the daily product TRMM version 7 3B42 provided by the Goddard Earth Sciences Data and Information Services Center (Mission, 2011) at https://disc.gsfc.nasa.gov/datasets/TRMM_3B42_7/.

A set of microwave and infra-red sensors onboard low earth orbit (LEO) satellites, such as the Microwave Imager (TMI) and the Advanced Microwave Scanning Radiometer-Earth Observing System (AMSR-E), provide the main source of information about hydrometeors for TRMM. The microwave sensor data is used to calibrate the infrared data to produce a first estimate of precipitation. However, even using the products of several satellites there is a sparse sampling of time-space precipitation in passive microwave techniques. Therefore, this data is complimented by infrared

Table 3.1: Summary of the datasets used in this study. For each dataset, the acronym used hereafter, the period of coverage, the field used and the horizontal resolution are shown. Some datasets extend further back in time, but only the satellite-era period is used in most of the datasets. The variables used are: precipitation, surface-air temperature ($2mT$), sea-level pressure (SLP), SSTs, the x and y components of the wind (u, v), the lagrangian tendency of air pressure (ω), outgoing longwave radiation (OLR), geopotential height (GPH) and specific humidity (q).

<i>Dataset/ Version</i>	<i>Acronym</i>	<i>Variable</i>	<i>Period</i>	<i>Data type</i>	<i>Resolution</i>	<i>Reference</i>
Global Precipitation Climatology Project v2.3	GPCP	Precipitation	(1979-2018)	Surface and satellite	2.5°x2.5°	(Adler et al., 2003)
Global Precipitation Climatology Centre	GPCC	Precipitation	(1940-2013)	Surface station	0.5°x0.5°	(Becker et al., 2011)
Climate Prediction Center	CMAP	Precipitation	(1979-2016)	Satellite calibrated with surface rain-gauge	2.5x2.5°	(Xie and Arkin, 1997)
Merged Analysis of Precipitation	CRU4	Surface temperature	(1979-2017)	Surface station	0.5°x0.5°	(Harris et al., 2014)
Climatic Research Unit TS v4.	CHIRPS	Precipitation	(1981-2018)	Surface rain-gauge and satellite	0.05°x0.05°	(Funk et al., 2015)
Climate Hazards Infrared Precipitation with Stations	TRMM	Precipitation	(1999-2018)	Satellite calibrated with surface station	0.25°x0.25°	(Huffman et al., 2010)
Tropical Rainfall Measurement Mission 3B42 V7	HadSST	SST	(1940-2018)	Buoy and satellite	2.5°x2.5°	(Kennedy et al., 2011)
Hadley Centre SST3	ERA-5	$2mT$, SLP, u , v , ω , OLR, q , SST, GPH, precipitation	(1979-2018)	Reanalysis	0.75x0.75°	(C3S, 2017; Hersbach et al., 2020)

measurements onboard geosynchronous earth orbit satellites. Other sources of information include a radar onboard TRMM and rain gauge analysis. Details of the research product can be found in Huffman et al. (2007) and Huffman et al. (2010).

The Climate Prediction Center Merged Analysis of Precipitation (CMAP) dataset is a global merged product of satellite and ground based observations but also constrained by a numerical model (Xie et al., 2007). This dataset was first produced at monthly-mean resolution (Xie and Arkin, 1997) but is now available as a collection of products at several temporal scales. The pentad-scale version of CMAP is used in this study.

The Climate Hazards Infrared Precipitation with Stations (CHIRPS) is relatively more recent merged product of precipitation (Funk et al., 2015). This dataset uses high-resolution rain-gauge station data that is complimented by satellite cloud cold duration estimates on regions where station data is sparse. The products are calibrated with TRMM data (Funk et al., 2015), so they are cannot be considered an independent source of information from TRMM.

All these datasets have shortcomings, advantages and unceratainties in their representation of precipitation. The algorithm of merged products such as TRMM to combine different satellite sensors and calibration techniques as well as surface station rain-gauge data results in products that may have shortcomings to accurately depict extreme events (Trejo et al., 2016), As the source data of most of these datasets is shared, the datasets cannot be considered to be fully independent sources of information.

The TRMM dataset has a high horizontal and temporal resolution and was used in several CMIP assessments (Geil et al., 2013; Jones and Carvalho, 2013) as a reliable source of precipitation (Carvalho et al., 2012). Therefore, TRMM is used in this thesis as the best estimate for the spatial and temporal characteristics of rainfall. However, the period covered by TRMM (1998-2018) is too short to analyse statistically robust teleconnections or variability, so GPCP, GPCC and CHIRPS are used to evaluate longer term variability for their longer period. Although a thorough validation and comparison of these datasets across the AMS domain is missing, several studies have analysed one or more of these datasets in regions of the AMS (e.g. Franchito et al., 2009; Dinku et al., 2010; Trejo et al., 2016).

3.1.2 ERA-5

A reanalysis is a numerical description of the state of the atmosphere in a global or regional scale of full of diagnostics in a gridded product that is available at multiple

vertical levels, in other words "maps without gaps" (Hersbach et al., 2020). A reanalysis takes a set of physically consistent blend of observations that are then used to run the forecasting model. Reanalysis are key tools for climate and weather research as they provide full pictures of the atmospheric state for long periods, a feature that could not be possible with our current purely observational tools. For this reason, reanalyses are typically used to validate GCM output.

The latest reanalysis from ECMWF, the fifth generation of their reanalysis, is called ERA5. ERA5 uses the Integrated Forecasting System (IFS) model and a 4D variational data assimilation system (4D-Var), a larger number of data sources for assimilation and also provides output on higher horizontal resolution (Hersbach et al., 2020). The output is available on hourly to monthly-mean frequencies, from 1000 hPa to 0.1 hPa in the vertical and with horizontal resolutions as high as 0.25° . In this thesis, the resolution of all ERA5 data, unless otherwise stated is of 0.75° and all data was downloaded from the Climate Store at <https://climate.copernicus.eu/climate-reanalysis>.

ERA5 presents a notable improvement in the representation of the water cycle, by increasing the mean correlation to precipitation datasets such as GPCP. ERA5, as all reanalysis, resolves precipitation rates in the driving physical model using the convective scheme and parametrisation. However, this reanalysis also assimilates radiances from several satellite instruments such as Global precipitation monitoring mission Microwave Imager, TMI and ASMR-2 Hersbach et al. (2020). This assimilation of satellite data has improved the representation of the water cycle in ERA5 compared to older reanalysis such as ERA-interim (e.g. Hénin et al., 2018). These improvements are also due to changes to the parametrisations of the microphysics of clouds and rain (Forbes and Ahlgrimm, 2014) and the diurnal cycle of convection (Bechtold et al., 2014).

3.2 The Unified Met Office Hadley Centre Model

The UK Met Office Hadley Centre (MOHC) has developed coupled climate models since 1999, releasing the first version of the Hadley Centre Global Environmental Model (HadGEM1) in 2006 (Johns et al., 2006). The HadGEM model has been updated continuously and is usually part of the Coupled Model Intercomparison Project (CMIP) experiments. This section first describes the third generation of the HadGEM model and subsequently describes the set up of experiments for CMIP6.

3.2.1 The Global Coupled Configurations of HadGEM3

The MOHC Unified Model (UM) is a weather and climate global model; based on a seamless modelling approach, the UM consists of a dynamical core and parametrization schemes that can be used across a wide range of temporal and spatial scales. The UM version that is used in throughout this thesis employs the Global Coupled (GC) configuration 3.1 (GC3.1) (Williams et al., 2018; Walters et al., 2019) which in turn is composed of the components: Global Atmosphere 7.0 (GA7.0), Global Land 7.0 (GL7.0), Global Ocean 6.0 (GO6.0), and Global Sea Ice 8.0 (GSI8.0). The GC3.1 configuration runs with 85 atmospheric levels, 4 soil levels and 75 ocean levels.

In most GCMs precipitation is simulated via two pathways. First, precipitation due to grid-scale processes, i.e., large-scale precipitation, is estimated by the micro-physics – or cloud – scheme that evaluates the cloud fraction and saturation within the grid-box column where precipitation occurs by processes such as warm rain (Walters et al., 2019). In turn, precipitation associated with convection is typically a sub-grid process and is calculated by the convection scheme. In GC3.1 the convective scheme follows three stages, according to Walters et al. (2019), first a diagnosis of the boundary layer to establish whether convection should occur at a given time-step and separately diagnosis shallow or deep convection, second, the shallow or deep convection schemes are called and third, a call for mid-level convection. In other words, the scheme first decides whether the thermodynamic profile at each grid-point fits certain parameters regarding buoyancy and vertical velocity, which then separates areas of deep and shallow convection, after these schemes the model implements the mid-level convection scheme to remove instabilities from above the shallow convective regions or below the deep convective regions. Further details of the GC3 configuration including model description and biases can be found in Williams et al. (2018) and Kuhlbrodt et al. (2018).

3.2.2 The CMIP6 experiments

The Met Office Hadley Centre (MOHC) has submitted the output of two models for CMIP6: HadGEM3 GC3.1 (hereafter GC3) is the latest version of the Global Coupled (GC) Met Office Unified Model (UM) and UKESM1, the new U.K. Earth System Model. The GC3 model was run for CMIP6 deck experiments with two horizontal resolutions: a low resolution configuration, labelled as N96, with an atmospheric resolution of $1.875^\circ \times 1.25^\circ$ and a 1° resolution in the ocean model and a medium

Table 3.2: Summary of the CMIP6 simulations in this study. For each simulation the acronym used hereafter, the experiment and the horizontal resolution are shown. The first 100 years of the piControl simulations are used and for historical experiments the period 1979-2014 is used.

Model	Experiment	Period	Atmospheric (Ocean) resolution	Acronym	Ensemble members	Reference
Hadley Centre Global Environment Model version 3 (HadGEM3)	Pre-industrial control	1850-2350	N96 1.875°x1.25° (1°)	GC3 N96-pi	1	(Menary et al., 2018; Ridley et al., 2018)
	Pre-industrial control	1850-2000	N216 0.83°x0.56° (0.25°)	GC3 N216-pi	1	(Menary et al., 2019c)
HadGEM3	Historical	1979-2014	N96 1.875°x1.25° (1°)	GC3-hist	4(r1-r4)	(Andrews et al., 2020; Ridley et al., 2019b)
	Historical	1979-2014	N216 0.83°x0.56° (0.25°)	N216-hist	1	(Ridley et al., 2019c)
HadGEM3	Atmospheric Model Intercomparison (AMIP)	1979-2014	N96 1.875°x1.25° (1°)	GC3-amip	5 (r1-r5)	(Ridley et al., 2019a)
	Pre-industrial control	2060-2600	N96 1.875°x1.25° (1°)	UKESM-pi	1	(Tang et al., 2019b)
United Kingdom Earth System Model version 1 (UKESM1)	Historical	1979-2014	N96 1.875°x1.25° (1°)	UKESM-hist	5 (r1-r5)	(Tang et al., 2019a)

resolution configuration, labelled N216, with atmospheric resolutions of $0.83^\circ \times 0.56^\circ$ and a 0.25° oceanic resolution (Menary et al., 2018).

The UKESM1 was recently developed aiming to improve the UM climate model adding processes of the Earth System (Sellar et al., 2019). These additional components include ocean biogeochemistry with coupled chemical cycles, tropospheric-stratospheric interactive chemistry which aim to better characterise aerosol-cloud and aerosol-radiation interactions (Mulcahy et al., 2018; Sellar et al., 2019). The physical atmosphere-land-ocean-sea-ice core of the HadGEM3 GC3.1 underpins the UKESM1, so that the UKESM1 and the HadGEM3 have the same dynamical core but the UKESM1 has the additional components mentioned above.

This study uses three CMIP6 deck experiments. First, the pre-industrial control (piControl) simulations, which are run with constant forcing using the best estimate for pre-industrial (1850) forcing of aerosols and greenhouse gas levels. The historical experiments are 164-yr integrations for 1850-2014 that include historical forcings of aerosol, greenhouse gas, volcanic and solar signals since 1850 (Eyring et al., 2016; Andrews et al., 2019). For further details, Andrews et al. (2020) extensively describes the historical simulations of HadGEM3-GC3.1.

In contrast to the pre-industrial control experiments, the historical experiments use time-varying aerosol and greenhouse gas emissions and land-use change (Eyring et al., 2016). In Latin-America, land-use change for agricultural purposes has dramatically decreased tree cover in Central America and south-eastern Brazil since the 1950s (Lawrence et al., 2012), thereby affecting the surface energy balance. The regional emissions of carbonaceous aerosols, nitrogen oxides and volatile organic compound in Latin America are also considered in the historical experiments. These emissions are noteworthy, e.g., due to the impact of black carbon emissions by increased biomass burning in the Amazon and northern Central America (Chuvieco et al., 2008).

The historical experiments of HadGEM3 and UKESM1 are composed of 4 and 9 ensemble members, respectively, but the results will be presented as the ensemble mean for the 1979-2014 period. These experiments will be referred to as GC3-hist and UKESM1-hist hereafter. Finally, we use the five ensemble members of the AMIP experiment from GC3 N96 covering 1979-2014. Table 3.2 summarises the main features of the experiments used in this study.

3.3 The Moist Static Energy Budget

The moist static energy (MSE) has been useful to diagnose changes to precipitation in observed () and simulated environments (). The MSE is given by:

$$h = L * q + C_p T + gz = Lq + s \quad (3.1)$$

where h denotes the MSE, C_p is the heat capacity at constant pressure, T is the air temperature, L is the latent heat of vaporization, q is the specific humidity and s is the dry static energy.

Chapter 4

The American monsoon system in UKESM1 and HadGEM3

This chapter evaluates the representation of in two state-of-the-art climate models: UKESM1 and HadGEM3. The models are assessed in three deck experiments submitted to CMIP6: the pre-industrial control, historical and atmosphere-only experiments described in section 3.2. The simulations show a good representation of the seasonal cycle of temperature in monsoon regions, although the historical experiments overestimate the observed summer temperature in the Amazon, Mexico and Central America by more than 1.5 K. The seasonal cycle of rainfall and general characteristics of the North American Monsoon of all the simulations agree well with observations and reanalysis, showing a notable improvement from previous versions of the HadGEM model. The models reasonably simulate the bimodal regime of precipitation in southern Mexico, Central America and the Caribbean known as the midsummer drought, although with a stronger than observed difference between the two peaks of precipitation and the dry period. Austral summer biases in the modelled Atlantic Intertropical Convergence Zone (ITCZ), cloud cover and regional temperature patterns are significant and influence the simulated regional rainfall in the South American Monsoon. These biases lead to an overestimation of precipitation in southeastern Brazil and an underestimation of precipitation in the Amazon. The precipitation biases over the Amazon and southeastern Brazil are greatly reduced in the AMIP simulations, highlighting that the Atlantic SSTs are key for representing precipitation in the South American Monsoon. El Niño Southern Oscillation (ENSO) teleconnections, of precipitation and temperature, to the AMS are reasonably simulated by all the experiments. The precipitation responses to the positive and negative phase of ENSO in subtropical America are linear in both pre-industrial and historical experiments. Overall, the biases in UKESM1 and the low resolution configuration

of GC3 are very similar for precipitation, ITCZ and Walker circulation, i.e., the inclusion of Earth System processes appears to make no significant difference for the representation of the AMS rainfall. In contrast, the medium resolution HadGEM3 N216 simulation outperforms the low-resolution simulations due to improved SSTs and circulation.

4.1 Introduction

Climate research in recent decades has aimed to reduce uncertainty in climate projections by improving GCMs, but different approaches taken by modelling centres are seemingly disconnected (Jakob, 2014). One approach is to reduce horizontal grid spacing down to km resolution to rely less on parametrizations and more on physical laws to represent clouds and convection (Palmer and Stevens, 2019). A second approach aims to include new explicit representation of Earth System processes to better characterise complex land-atmosphere-ocean biogeochemical cycles that may provide a better constraint on climate sensitivity, a parameter that depends on the carbon cycle (Marotzke et al., 2017; Sellar et al., 2019; Andrews et al., 2019). Finally, recent modelling centres have chosen to include stochastic parametrisations of sub-grid processes since this approach has improved seasonal forecasts and may therefore improve climate projections (Palmer, 2019).

Model validation and assessment is important to analyse the effect of new parametrisations and to highlight missing processes but also evaluate which route provides the more substantial model improvement, stochastic parametrisations, increased resolution or Earth System processes. In this case, the focus of this chapter is to evaluate two models in the American monsoon system (AMS), one coupled ocean-atmosphere model HadGEM3 run at two horizontal resolutions compared to the Earth System model UKESM1, described in 3.2. Therefore, this chapter compares the effect of increased horizontal resolution and Earth System processes on the representation of the AMS in a climate model. In this chapter, in addition to the historical experiments which are more suitable for model assessment, both the pre-industrial control and atmosphere-only experiments are also used.

The assessment of climate models in monsoon regions is key to understand current and future changes to the water cycle in tropics. However, in the AMS, model assessments are usually only done in a handful of studies per CMIP phase. These studies only provide a wide view of the biases of each generation of models while usually highlighting which biases have improved and which biases remain from previous

model generations. However, a deeper evaluation of individual models can be used to provide better insight into the processes associated with climatological biases in key diagnostics such as temperature and precipitation.

For example, in the South American Monsoon, CMIP5 models improved from CMIP3 in the simulated distribution of precipitation during monsoon maturity and exhibited an improved seasonal cycle (Jones and Carvalho, 2013; Yin et al., 2013). However, long-term biases in the South American Monsoon, e.g., the underestimation of rainfall in the central Amazon, persist in CMIP5 (Yin et al., 2013). The geographic distribution of rainfall during austral fall and several characteristics of the South Atlantic Convergence Zone are also poorly represented in CMIP5. However, these studies provided little evidence as for the reasons for the improvements or the remaining biases in the models. A clear motivation to evaluate models in the South American Monsoon is that the accurate simulation of the geographic distribution and seasonality of rainfall in the Amazon rainforest is a relevant issue due to the impact of the rainforest on climate and society (e.g. Li et al., 2006; Malhi et al., 2009; Yin et al., 2013) and thus more research on the representation of the South American Monsoon is warranted.

The remainder of this chapter is organised as follows, section 4.2 compares modelled and observed climatological temperature, sea-level pressure and low-level wind fields, whereas section 4.3 analyses the Pacific and Atlantic ITCZs. Section 4.4 analyses the spatial and temporal characteristics of rainfall and convection in the AMS while section 4.5 documents the simulated teleconnections of ENSO. A summary and discussion of the results is provided at the end of the chapter.

4.2 Climatological temperature and low-level wind biases

This section evaluates the simulated climatological temperature and low-level wind structure in the AMS region. The climatological representation of the near-surface air temperature and low-level winds in the models is compared to ERA5 in Figures 4.1 and 4.2, the climatology of DJF and JJA of ERA5 is shown in Figure 4.1a, b. The biases of the historical experiments, computed as the differences between the model and observed fields, are shown in Figures 4.1c, d) for GC3-hist and e, f) for UKESM1-hist. Only statistically significant differences are shown, according to a Welch t-test (Wilks, 2011), which accounts for the difference in sample size and variance between model and observations/reanalysis data. The significance for

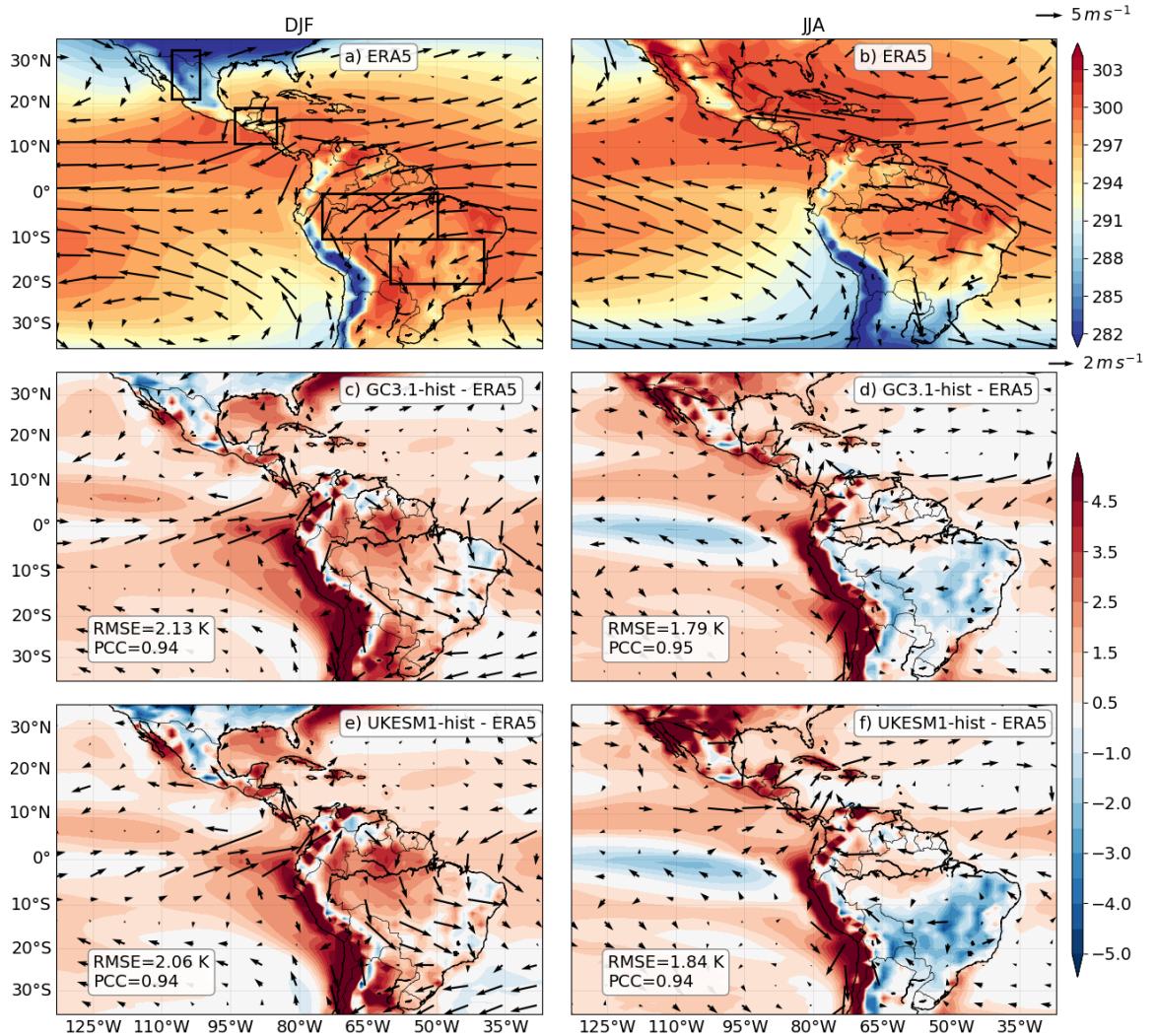


Figure 4.1: (a, b) Temperature (color-contours in K) and wind speed (vectors) at 850 hPa DJF and JJA climatologies in ERA5. The biases are shown as the differences between the ensemble mean from the historical experiment of (c, d) GC3 and (e, f) UKESM1 and ERA5. The climatogies and biases are shown for (a, c, e) boreal winter (DJF) and (b, d, f) boreal summer (JJA). Only differences statistically significant to the 95% level are shown, according to a Welch t-test for each field. The key for the size of the wind vectors is shown in the top right corner of panels b) and d). The root-mean square error (RMSE) and pattern correlation coefficient (PCC) are shown on the bottom left of c-f.

simulations with multiple ensemble members is estimated first for each ensemble member and then combined into a single probability or p-value using Fisher's method (Fisher, 1992). Pattern correlations and root-mean square error (RMSE) are shown in Figures 4.1c-f and in Table 4.1.

During DJF, the simulations show a colder-than-observed sub-tropical North Amer-

Table 4.1: Root-mean square error (RMSE) and pattern correlation coefficients (PCC) for each season and each model experiment. Near surface air temperature ($t2m$), wind components (u and v) and mean-sea level pressure ($mslp$) are assessed against ERA-5 and precipitation (pr) against TRMM.

<i>Variable</i>	<i>Model experiment</i>	DJF RMSE	DJF PCC	MAM RMSE	MAM PCC	JJA RMSE	JJA PCC	SON RMSE	SON PCC
t2m	GC3 N96	1.28	0.98	1.3	0.96	1.38	0.96	1.31	0.96
t2m	GC3 N216	1.05	0.99	1.07	0.98	1.02	0.98	0.98	0.98
t2m	GC3 Hist	2.06	0.94	1.75	0.93	1.73	0.94	2.05	0.92
t2m	UKESM-hist	2.03	0.94	1.77	0.93	1.8	0.94	2.0	0.93
t2m	GC3 AMIP	1.17	0.98	1.12	0.97	1.2	0.97	1.2	0.97
u	GC3 N96	0.78	0.99	0.59	0.99	0.9	0.98	0.87	0.98
u	GC3 N216	0.78	0.99	0.59	0.99	0.9	0.98	0.87	0.98
u	GC3 Hist	1.02	0.98	1.04	0.97	0.92	0.98	0.84	0.98
u	UKESM-hist	1.04	0.98	1.01	0.97	0.91	0.98	0.82	0.98
u	GC3 AMIP	0.96	0.98	0.77	0.99	1.18	0.97	1.09	0.96
v	GC3 N96	0.75	0.93	0.66	0.93	0.65	0.95	0.59	0.94
v	GC3 N216	0.6	0.96	0.5	0.95	0.57	0.96	0.54	0.94
v	GC3 Hist	0.76	0.94	0.72	0.92	0.66	0.95	0.59	0.94
v	UKESM-hist	0.75	0.93	0.69	0.92	0.65	0.95	0.6	0.93
v	GC3 AMIP	0.67	0.95	0.52	0.95	0.68	0.94	0.61	0.93
mslp	GC3 N96	1.33	0.96	1.03	0.97	1.15	0.96	0.95	0.97
mslp	GC3 N216	1.11	0.97	0.9	0.97	1.1	0.96	0.89	0.97
mslp	GC3 Hist	1.31	0.97	1.12	0.96	1.08	0.96	0.94	0.97
mslp	UKESM-hist	1.4	0.97	1.15	0.96	1.14	0.95	0.99	0.97
mslp	GC3 AMIP	1.15	0.97	0.87	0.97	1.09	0.96	0.93	0.97
pr	GC3 N96	2.02	0.79	2.24	0.71	1.62	0.9	1.69	0.86
pr	GC3 N216	1.58	0.88	1.72	0.85	1.4	0.93	1.57	0.89
pr	GC3 Hist	2.05	0.78	2.49	0.64	1.69	0.88	1.69	0.86
pr	UKESM-hist	1.96	0.8	2.39	0.66	1.71	0.88	1.62	0.87
pr	GC3 AMIP	1.42	0.9	1.61	0.88	1.95	0.88	1.8	0.88

ica and a warm bias over the Amazon ($\approx +3.5$ K). The west coast of South America also shows a significant warm bias ($> +4$ K) in the historical simulations. The simulated circulation in austral summer in South America has a significant bias in the easterly flow coming from the equatorial and subtropical Atlantic. The low-level wind biases suggest a weaker easterly flow from the Atlantic into southeastern Brazil but also a strong southward flow from northern to southern South America. The South America Low-Level Jet, the low-level northwesterly flow in Bolivia, observed in Figure 1a, is stronger in the simulations. This stronger than observed jet is suggestive of a stronger moisture transport to the La Plata Basin, which has been associated with a drying of the Amazon and positive precipitation anomalies at the exit region of the jet (Marengo et al., 2012; Jones and Carvalho, 2018).

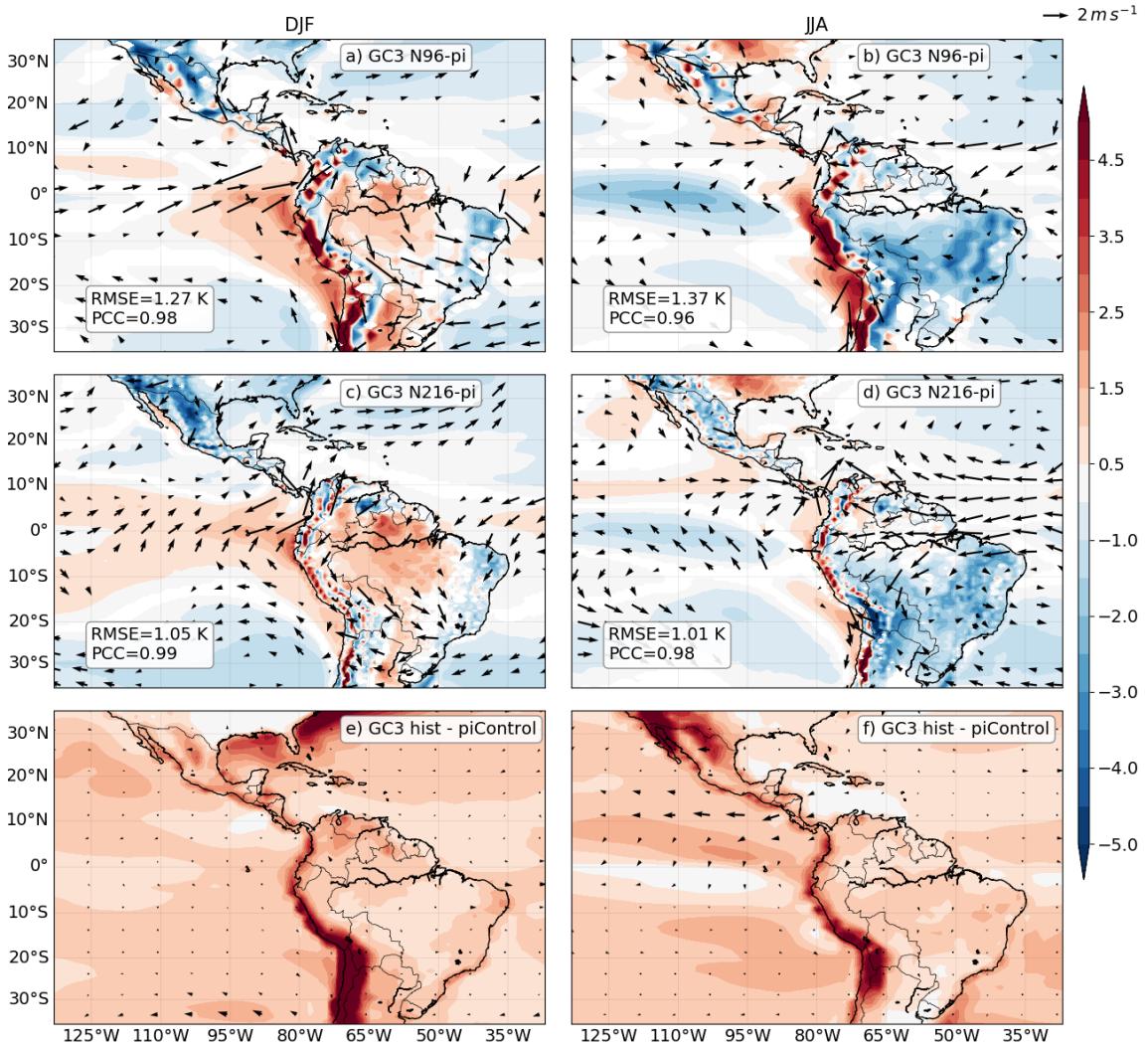


Figure 4.2: As in Figure 4.1, but showing the differences between the piControl simulations of (a, b) GC3 N96-pi and (c, d) GC3 N216-pi, and ERA5. (e, f) show the statistically significant differences between the historical (1979-2014) and piControl experiments of GC3. The RMSE and PCC are shown on the bottom left of a-d.

In turn, in boreal summer (Figures 4.1d, f), positive temperature biases are observed in southwestern North America ($> +3.5$ K), which are higher in UKESM1-hist than in GC3-hist. The easterly flow west of Central America has a negative bias in UKESM1 suggesting a weaker flow that crosses from the Caribbean Sea into the East Pacific Ocean. Also in JJA, the simulated East Pacific surface temperatures are colder than observed for both historical experiments. The inclusion of Earth System processes appears to make no improvement on the low-level circulation biases.

The piControl simulations (Figures 4.2a-d) have some similar biases to the historical simulations. In DJF, the piControl simulations show a similar but smaller

positive bias in the Amazon than the historical experiments, as well as a similar bias in the circulation in South America, with the smallest biases in GC3 N216 piControl. In JJA, the piControl simulations do not show the positive temperature bias in north-western North America. However, the bias in the zonal wind over the easternmost Pacific is present in both piControl and historical simulations.

Figures 4.2e, f show the difference between the historical and piControl experiment of GC3, illustrating the response to historical forcing in GC3. The temperature response in austral summer in South America is observed as 1.5 K whereas in JJA in North America temperatures were 4 K higher in the historical experiment than in the piControl. A very similar temperature pattern response to historical forcing was observed for UKESM1 (not shown) although of slightly different magnitude. The only significant difference in low-level winds, as a response to historical forcing, are the easterlies in the East Pacific Ocean during JJA, which are stronger in the historical simulation.

The seasonal cycle of temperature in key regions (depicted in Figure 4.1a) of the AMS is shown in Figure 4.3, comparing the simulations to ERA5 and the CRU4 dataset. The temperature in the North American Monsoon region ranges from the boreal winter 12°C to a maximum in June close to 27°C. Although the piControl simulated temperatures are colder than observed throughout the year, the models reasonably reproduce the seasonal cycle, which may be relevant for the simulated monsoon onset timing and strength (Turrent and Cavazos, 2009). The historical experiments notably show a colder than observed winter and a warmer than observed summer.

The piControl simulations show a colder-than-observed winter in southern Mexico and northern Central America. The historical experiments show a warming signal, when compared to the piControl simulations, are warmer by about 1.5 K in winter and 2 K in the summer. In spite of these biases, both types of experiments follow closely the seasonal cycle in North and Central America.

However, the seasonal cycle in South America is shown in Figures 4.3 c, d for southeastern Brazil and the central Amazon. The simulations show a stronger than observed seasonal cycle, especially the historical experiments. For example, the modelled temperature difference between late austral winter and spring was \approx 4 K whereas the observed temperature varies by less than 1 K in the same period. The models show a warm bias in the Amazon region (Fig. 4.3 d) which peaks in austral spring (SON), during the development of the monsoon (Marengo et al., 2012). In south-eastern Brazil, the seasonal cycle is reasonably well reproduced but with a significant

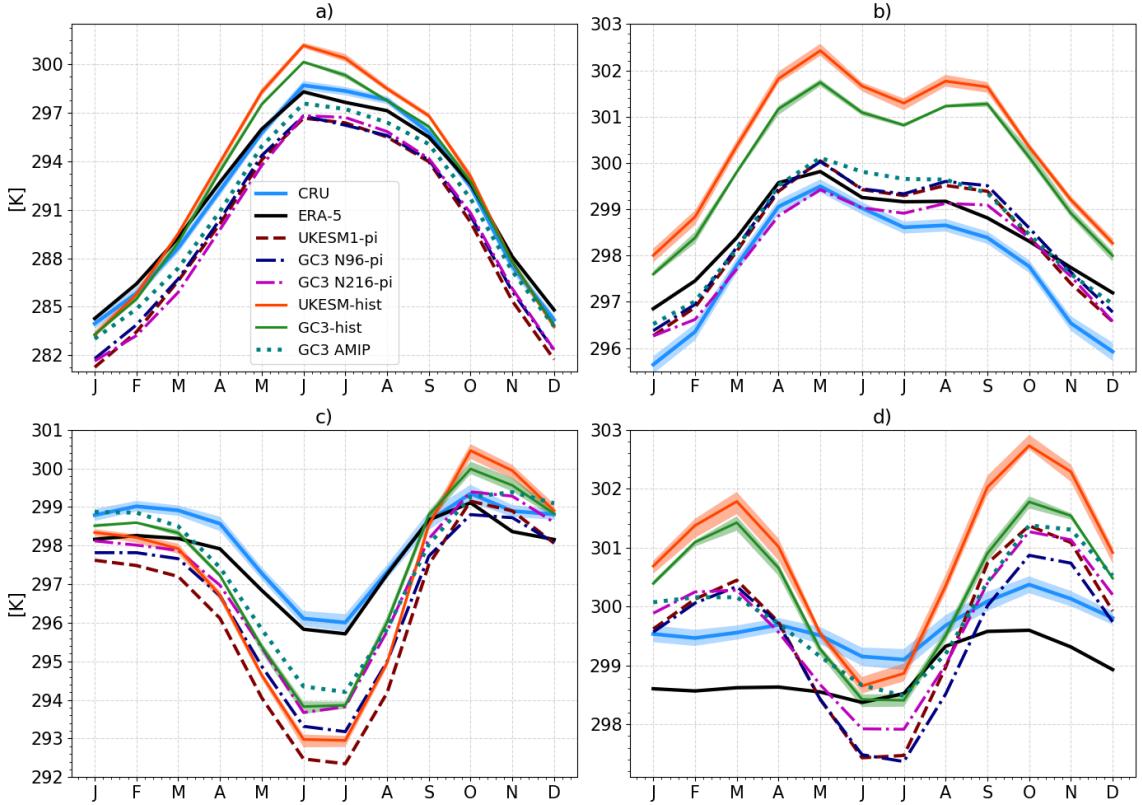


Figure 4.3: Monthly-mean temperature in the (a) North American Monsoon [$19\text{--}35^\circ\text{N}, 110\text{--}103^\circ\text{W}$], (b) the Midsummer drought [$11\text{--}19^\circ\text{N}, 95\text{--}85^\circ\text{W}$] (c) Eastern Brazil [$20\text{--}10^\circ\text{S}, 60\text{--}40^\circ\text{W}$] and (d) the Amazon basin [$-10\text{--}0^\circ\text{S}, 75\text{--}50^\circ\text{W}$] regions. The shadings for the CRU dataset represents the observational uncertainties and for the historical simulations the shading is the ensemble spread. The regions for this plot are shown in Figure 4.1a.

cold bias throughout the year which maximizes during austral winter (JJA), as models (e.g. UKESM1) simulate a temperature 4 K lower than observed. In all panels of Figure 4.3, the historical experiments show a significant warming signal as a response to historical forcing, which is generally stronger in UKESM1 than in GC3.

4.3 The Atlantic and Pacific ITCZs and the SACZ

The AMS is intertwined with the seasonal migration of the East Pacific and Atlantic ITCZ as the ITCZ largely determines regions of ascending and descending motions, moisture transport and the hemispheric energy balance (Oueslati and Bellon, 2013; Li and Xie, 2014; Zhou et al., 2016; Cai et al., 2019). In particular, the North American monsoon and MSD are mostly influenced by the East Pacific ITCZ whereas the South

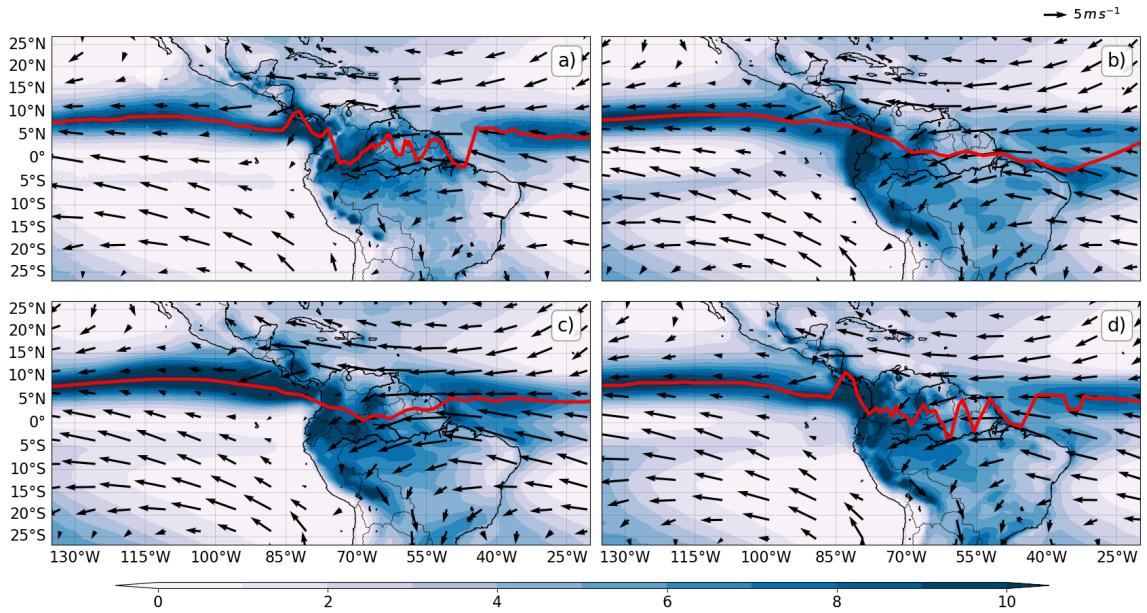


Figure 4.4: Climatological rainfall [mm day^{-1}] and low-level wind speed (850-hPa) in (a) TRMM and ERA-5, (b) the ensemble-mean UKESM-historical, (c) GC3-amip and (d) GC3 N216-pi. The red line highlights the maximum rainfall for each longitude as a proxy for the position of the ITCZ.

American monsoon is affected by the strength and position of the Atlantic ITCZ.

Figure 4.4 shows the observed and modelled climatological rainfall and the ITCZ climatological position. Three simulations are shown: the ensemble-mean UKESM1-historical, the ensemble mean GC3 AMIP and GC3 N216-pi. Other simulations are not shown as all the coupled low resolution simulations showed very similar precipitation and ITCZ characteristics whereas the AMIP and medium-resolution experiments showed notable differences.

The climatological ITCZ in TRMM (Figure 4.4a) is found, on average, at 8°N in the East Pacific and at 6°N in the Atlantic. All the simulations reasonably represent the climatological position of the East Pacific (EP) ITCZ; however, the modelled Atlantic ITCZ near the coast of Brazil is found south of the equator at 3°S in the coupled model simulations. The location of the ITCZ in GC3 N216-pi and the spatial distribution of rainfall is more consistent with TRMM dataset than the rest of experiments. Rainfall near the Amazon river mouth is significantly larger in the low resolution simulations than in the TRMM dataset. However, the GC3 AMIP shows the best agreement with TRMM in ITCZ position and rainfall distribution.

The seasonal cycle of the ITCZ location, precipitation rates and low-level winds in both basins are shown in Figure 4.5, for TRMM, UKESM1-hist, GC3 AMIP, GC3

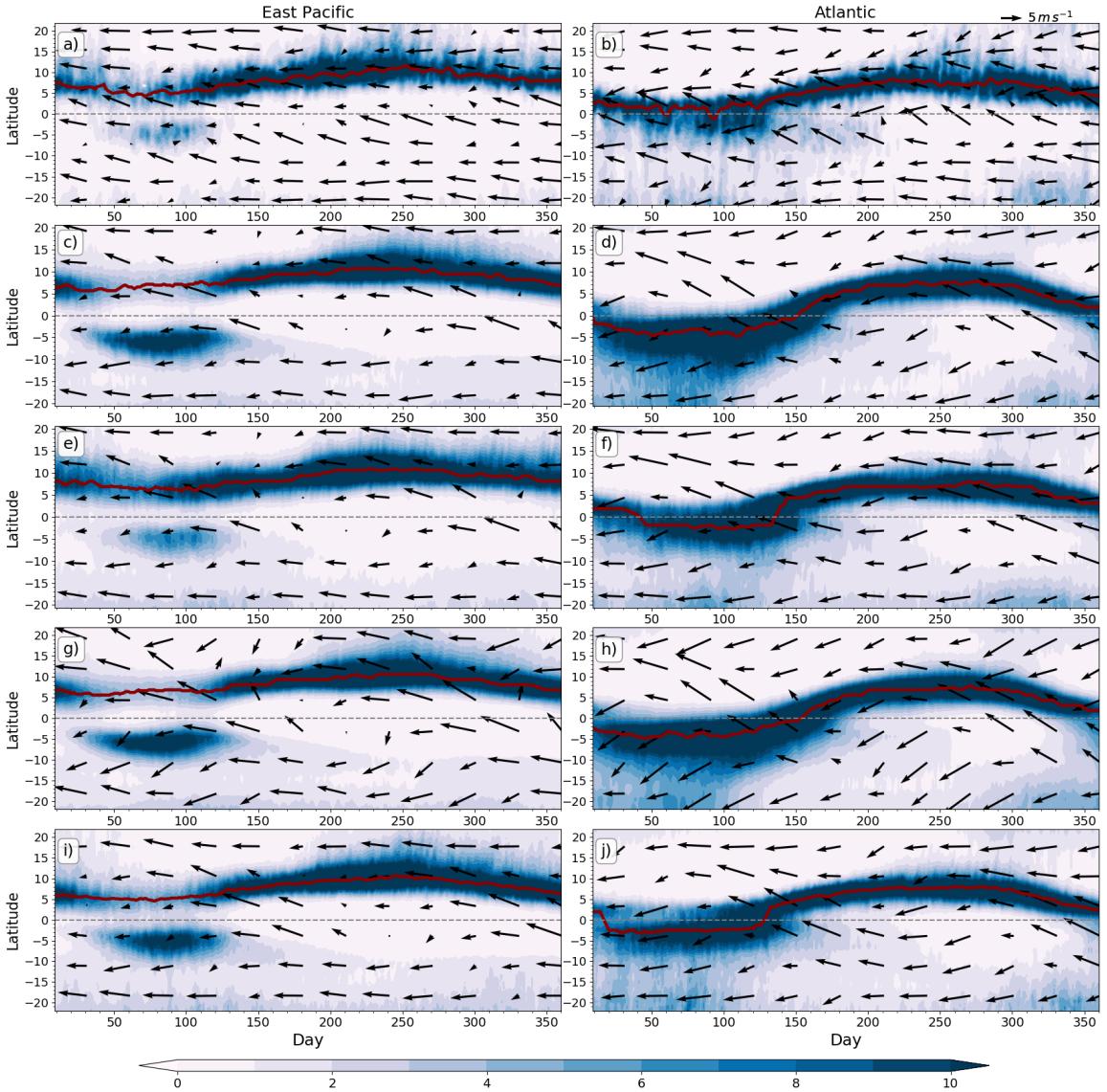


Figure 4.5: Time-Latitude plot of daily mean rainfall (colour contours) and low-level wind speed (850 hPa) longitudinally averaged over the (a, c, e, g) East Pacific [150°W - 100°W] and (b, d, f, h) Atlantic [40°W - 20°W] Oceans. (a, b) show rainfall from TRMM and winds from ERA-5, (c, d) the ensemble-mean UKESM-historical, (e, f) GC3 AMIP, (g, h) N96-pi and (i, j) GC3 N216-pi. The red solid line shows the ITCZ as the latitude of maximum precipitation.

N96-pi and GC3 N216-pi. The East Pacific (EP) ITCZ in observations (Fig. 4.5a) migrates southwards during the first days of the year and is weakest and at its southernmost position at 5°N around day 100 (mid-April). During boreal spring, the EP ITCZ migrates northward reaching a peak latitude and maximum rainfall at 10°N by day 250, or early September. The low-level winds are predominantly easterly, which

are stronger away from the ITCZ and weaker and convergent near the ITCZ position. The position and seasonal migration of the EP ITCZ is reasonably well represented in the four simulations (Figs. 4.5c, e, g, i), but a noticeable bias in precipitation is observed in boreal winter south of the equator in the coupled simulations. The modelled low-level winds in the coupled simulations show significant biases near the ITCZ. These wind biases are observed as stronger wind vectors converging toward the ITCZ during boreal summer and spring and stronger wind vectors diverging away from the equator during boreal winter.

The observed Atlantic ITCZ (Figure 4.5b) has a similar seasonal cycle to the EP ITCZ. The Atlantic ITCZ is close to 4°N at day 1 and migrates southwards at the start of the year reaching its southernmost position at 0° at the end of March. During boreal spring, the Atlantic ITCZ migrates north, reaching 8°N at the start of boreal summer. The boreal winter position of the modelled ITCZ is displaced south with respect to the observations. The simulated ITCZ crosses south of the equator during boreal winter, as high rainfall rates above 12 mm day^{-1} covering the $10\text{S}-0$ region. After boreal spring, the modelled ITCZ crosses back north of the equator and matches the observed ITCZ reasonably well for boreal summer and fall. Low-level wind vectors near the Atlantic ITCZ (Figures 4.5f and h) suggest a simulated southerly bias north of the equator and a stronger northerly flow south of 10°S .

The biases in the Atlantic ITCZ can also be observed in the overturning circulation (not shown) and the associated Walker circulation as significant negative ω and q biases just north and south of equatorial South America indicative of weaker convective activity. The Atlantic Ocean shows a biased strong ascent south of the equator and a biased weak ascent north of the equator in the low resolution simulations. These biases in the Atlantic ITCZ and overturning circulations described above were found to be of similar magnitude in all the coupled model simulations run at lower resolution, both historical and piControl experiments, however, these biases improved in the medium resolution GC3 N216-pi and in the AMIP simulations (Figures 4.5f, j).

The South Atlantic Convergence Zone (SACZ) is a northwest-southeast oriented band of convection and is a prominent influence on the South American Monsoon mean and extreme rainfall (Carvalho et al., 2004; Marengo et al., 2012; Jorgetti et al., 2014). The SACZ is primarily characterized by convergence oriented northwest-to-southeast that promotes rainfall in southeastern Brazil. The position of the SACZ and strength are an important factor for variability of the South American monsoon

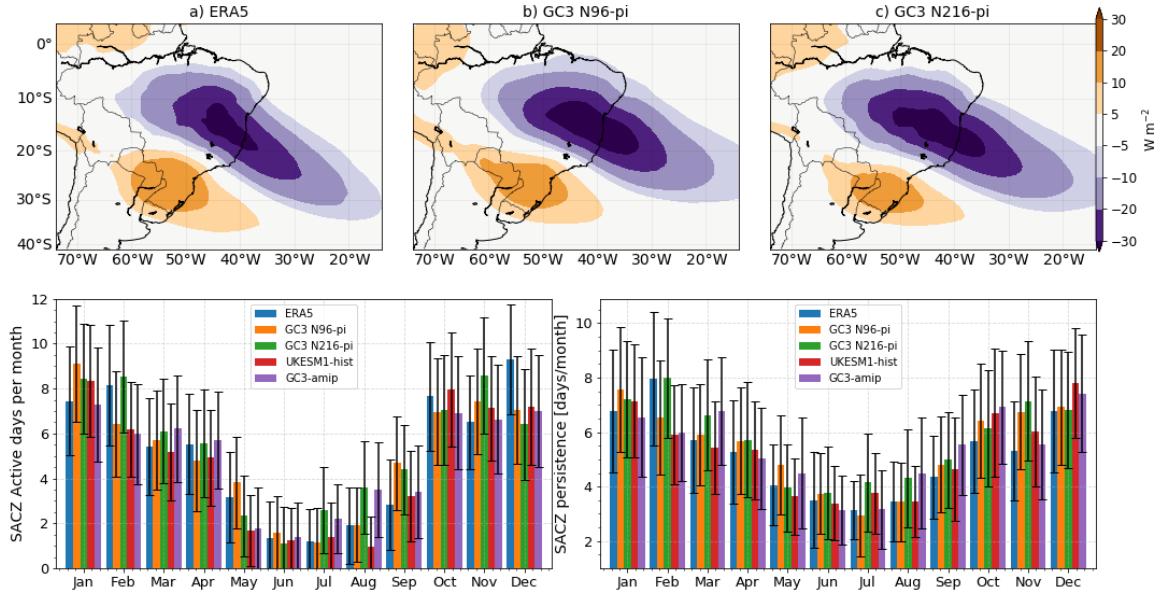


Figure 4.6: (a, b, c) OLR anomalies during active South Atlantic Convergence Zone (SACZ) events. (d, e) Frequency of active SACZ days and length of active SACZ events in reanalysis and model data, the standard deviation is shown as the error bar. The SACZ active days are constructed by first computing the first EOF of the monthly-mean deseasonalized OLR and then the daily OLR, previously filtered to remove periods higher than 99 days, is projected on the EOF pattern to produce a time-series of pseudo-principal components. Active SACZ days are found when this time-series of pseudo-PCs is greater than 1, and the persistence is measured as the number of continuous days where the time-series is greater than 1.

on different temporal and spatial scales (Carvalho et al., 2004; Marengo et al., 2012; Jorgetti et al., 2014).

Figure 4.6 evaluates the SACZ in ERA5 compared to several simulations. The SACZ characteristics analysed are the outgoing-longwave radiation pattern, the SACZ active days and the persistence of the SACZ. The simulations from UKESM1, and GC3 N96 and N216 appear to reasonably simulate the spatial pattern of active SACZ days characterized by the low OLR in southeastern Brazil and higher OLR in the La Plata Basin. Similarly, the seasonal cycle of the frequency and persistence of SACZ active days is very well represented by the models with peak activity from November through January and very little activity during austral winter.

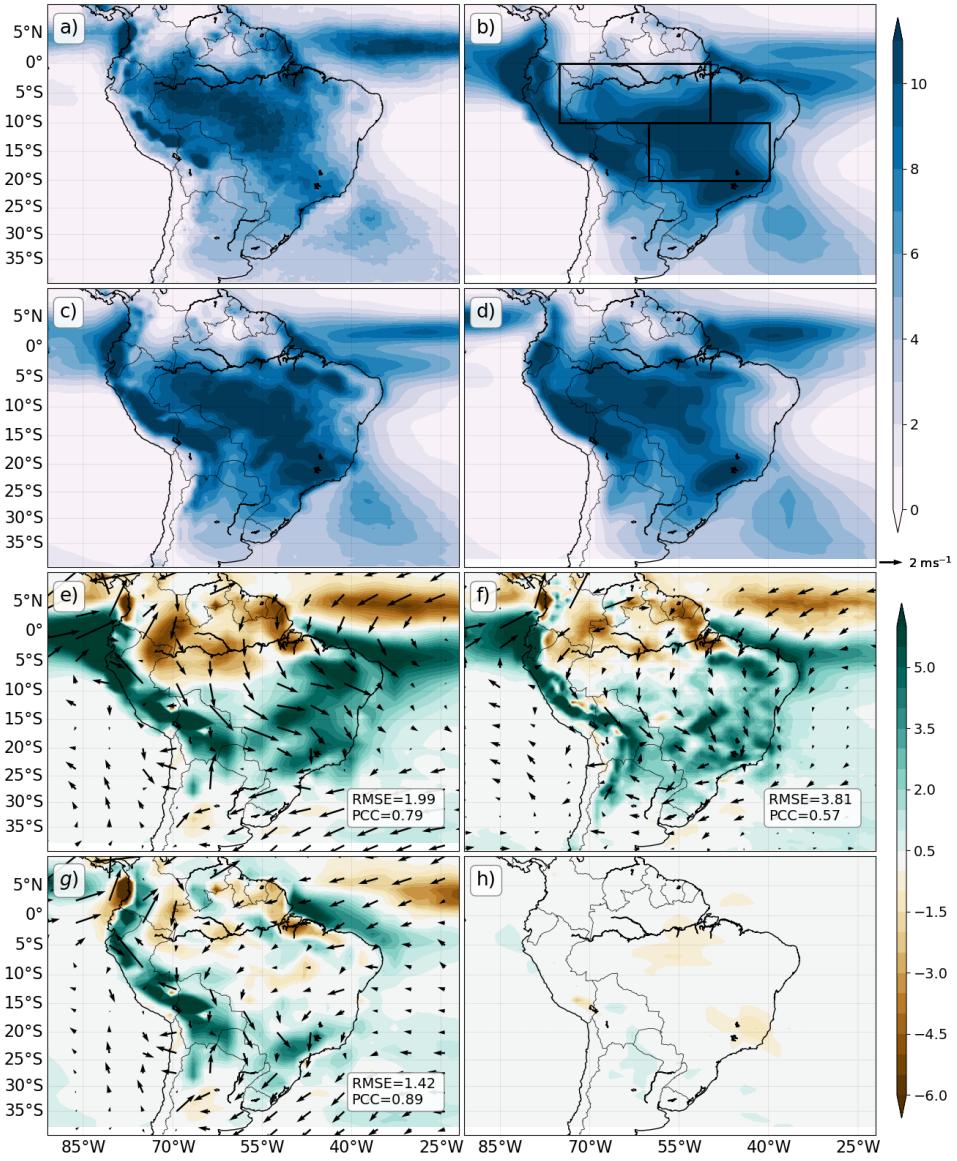


Figure 4.7: DJF mean rainfall [mm day^{-1}] from (a) TRMM, (b) UKESM1-historical, (c) GC3 N216-pi and (d) GC3-amip. (e, f, g) show the statistically significant differences between panels (b, c ,d) and (a) TRMM, respectively. (h) Precipitation difference between UKESM-historical and UKESM1-pi, only statistically significant differences (95%) confidence level is shown. The biases in the 850-hPa winds are shown as vectors.

4.4 Precipitation and convection in the AMS

4.4.1 Mean seasonal precipitation

The austral summer (DJF) rainfall distribution in South America is shown in Figure 4.7. The spatial distribution of rainfall is compared between the TRMM climatology

and the simulations for GC3 N216-pi, UKESM-hist and GC3-amip. The maximum austral summer rainfall in TRMM (Figure 4.7a) is found as a northwest-southeast oriented band of precipitation from the core Amazon region into southeastern Brazil, which is related to the SACZ. The biases are illustrated (Figures 4.7e-h) as the precipitation difference between the simulations and TRMM.

The coupled simulations show three main biases. Rainfall in the Atlantic ITCZ in these simulations is displaced southwards, observed as positive ($+5 \text{ mm day}^{-1}$) biases south of the equator and negative biases (-5 mm day^{-1}) north of the equator in the Atlantic. Second, the models underestimate rainfall in the core Amazon basin by -3 mm day^{-1} on average, and the third major bias is that rainfall in southeastern Brazil is overestimated by more than $+5 \text{ mm day}^{-1}$, approximately +100% of the observed rainfall in this region.

The precipitation biases are associated with a stronger northerly flow in South America, transporting moisture from the Amazon into southeastern Brazil and the La Plata Basin. The magnitude of these biases is smaller in GC3 N216 (Figure 4.7f) than in the low resolution simulations, such as UKESM1-hist. The ensemble mean GC3 AMIP (Figure 4.7d) shows a better representation of the austral summer rainfall and circulation patterns, removing the main biases (Figure 4.7g) of the coupled simulations. The response to historical forcing, illustrated by the difference between UKESM1-hist and UKESM1-pi (Figure 4.7h), is much weaker than the magnitude of the biases.

The modelled and observed JJA mean rainfall and biases for Mexico and Central America are shown in Figure 4.8. The main feature is the East Pacific (EP) ITCZ which extends north to 15°N near the western coast of Mexico as a broad band of rainfall ($\sim 11 \text{ mm day}^{-1}$). The modelled EP ITCZ (Figures 4.8e, f, g) rainfall is overestimated by more than 5 mm day^{-1} , especially in GC3-amip. This wet bias is associated with a westerly bias in the low-level circulation, suggesting a weaker flow from the Caribbean into the East Pacific.

The North American Monsoon can be observed as a band of precipitation across western Mexico. In the core monsoon region, near the Sierra Madre Occidental (Adams and Comrie, 1997; Zhou et al., 2016), the JJA-mean rainfall is higher than 6 mm day^{-1} . The distribution of rainfall in the North American Monsoon region is relatively well represented in all the simulations, as only a small wet bias ($+2 \text{ mm day}^{-1}$) in western Mexico is observed. The northernmost part of the North American Monsoon (southwestern US) is best simulated by GC3 N216-pi, as the other simulations show a dry bias in this region. The low-resolution simulations (Figure

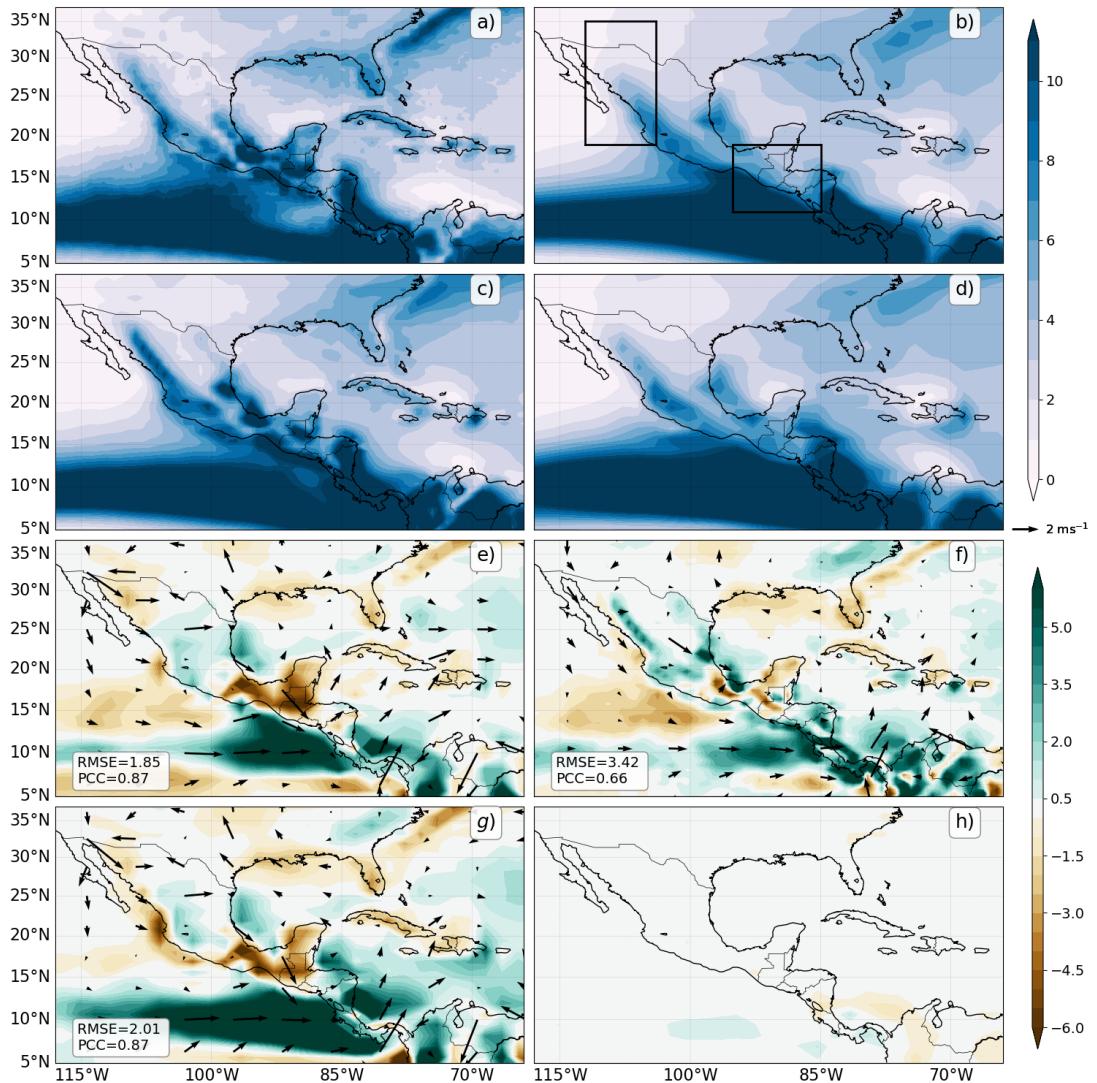


Figure 4.8: As in Figure 4.7 but for JJA in the northern part of subtropical America.

4.8e) underestimate rainfall (-5 mm day^{-1}) over land in southern Mexico, Guatemala and Belize. Rainfall in the Caribbean islands and Florida is underestimated (-1 mm day^{-1}) in all simulations.

In most cases for JJA in this region, the precipitation and wind biases were reduced in the high-resolution simulation (Figure 4.8f). The precipitation response to historical forcing is much lower than the biases (Figure 4.8h) with no significant precipitation differences over land due to the historical forcing.

4.4.2 The annual cycle of rainfall

Figure 4.9 shows the seasonal cycle of rainfall at the pentad (5-day) scale over the North American Monsoon, the Midsummer drought (MSD), the Amazon and eastern Brazil regions. The correlation between TRMM and the model and reanalysis data (ERA5) is also shown in each panel.

The seasonal cycle of precipitation in the MSD region in the simulations is well represented as all the simulations show the characteristic bimodal distribution, a feature that is uncommon for a climate model to be able to reproduce (Ryu and

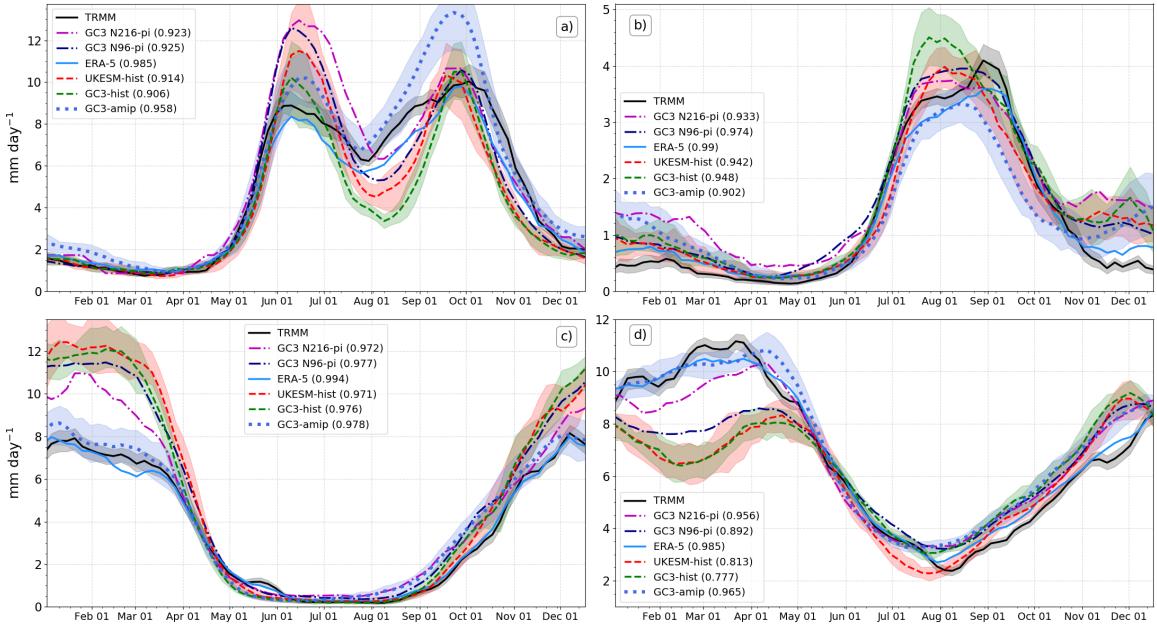


Figure 4.9: Annual cycle of pentad-mean rainfall in the regions (a) the Midsummer drought, (b) the North American Monsoon, (c) Eastern Brazil and (d) the Amazon Basin. The regions are defined as in Figure 4.3 and are illustrated in Figure 4.8b and Figure 4.9b. The shaded regions represent observational uncertainty for TRMM and ensemble spread for the historical experiments. The correlation coefficient for each of the simulated seasonal cycles with TRMM is given in brackets in each panel.

Hayhoe, 2014). However, the characteristics of the simulated MSD are different from observations. For example, the magnitude of the first peak and second peaks in the simulations are different. For instance, most of the first peak simulated magnitudes are higher than TRMM by 4 mm day^{-1} , and the AMIP simulation overestimates the second maximum of rainfall by $2\text{-}3 \text{ mm day}^{-1}$. Similarly, the differences between the first peak and the MSD and between the MSD and the second peak are more pronounced in the coupled simulations. The timing of the MSD period is different in the models, as the simulations show the driest period taking place 10 days after TRMM and ERA5.

In the North American Monsoon (Figure 4.9b), the observed seasonal cycle is characterized by a very long and dry period ranging from the end of November to the start of June, which is followed by a sharp increase of rainfall around mid-June. The timing of the increase of rainfall in models coincides with observations, suggesting that onset timing and strength is well represented in these models. Moreover, the modelled and observed mean precipitation rates during monsoon maturity are 4 mm day^{-1} , from mid-July until early September, which suggests notable ability of the models to reproduce the peak monsoon rainfall. The historical simulations show a shorter wet season characterised by an earlier retreat of the monsoon rainfall and, as all the simulations, a positive boreal fall rainfall bias ($+1 \text{ mm day}^{-1}$), a feature that has been shown in these models in CMIP5 (Geil et al., 2013) and that will be further explored in the following chapter.

The seasonal cycle of precipitation in eastern Brazil is characterised by a very wet summer ($\sim 8 \text{ mm day}^{-1}$) compared to a very dry ($\sim 0.2 \text{ mm day}^{-1}$) winter (Figure 4.9c). Austral summer rainfall in TRMM and ERA5 increases steadily from September to a maximum found in early January ($\sim 8 \text{ mm day}^{-1}$). Rainfall in this region decreases to $\sim 6 \text{ mm day}^{-1}$ by late March as the monsoon migrates northward and then sharply decreases in austral fall.

The models (Figure 4.9c) show a positive bias during monsoon maturity. This bias was found to be of $+4 \text{ mm day}^{-1}$ and $+2.5 \text{ mm day}^{-1}$ for the low and medium resolution simulations, respectively. This positive bias in the maximum rainfall is consistent with the biases shown in Figure 4.7, which showed that rainfall in south-eastern Brazil is overestimated, especially in the low resolution coupled simulations. In contrast to the coupled simulations, GC3-amip shows a very good agreement with the observed maximum summer rainfall and the seasonal cycle ($r=0.978$) throughout the year.

Finally, the seasonal cycle in the Amazon (Figure 4.9d) has a weaker contrast as rainfall is found year-round. The coupled simulations show a dry bias during austral summer and a good agreement with the observations during austral winter. Rainfall rates in the Amazon from January to March, in both TRMM and ERA-5, is close to 10 mm day^{-1} , yet the low resolution simulations show rainfall rates of 8 mm day^{-1} in mid-February, particularly the historical experiments. GC3 N216-pi shows a better agreement with observations but still underestimates summertime rainfall by 1 mm day^{-1} .

This dry Amazon bias has been a known feature of GCMs, including the MOHC models since CMIP3 (Li et al., 2006; Yin et al., 2013). In these simulations the dry Amazon bias is only alleviated in GC3-amip whose seasonal cycle and maximum summer rainfall agree well with observations. The models, however, represent with reasonable skill the transition from early austral spring (4 mm day^{-1} in September) to summertime rainfall (6 mm day^{-1} in November).

4.4.3 Characteristics of convective activity

The seasonal cycles of out-going longwave radiation (OLR), vertical velocity (ω) and specific humidity (q) are key features of a monsoon since these quantities characterise the strength and height of deep convection. Figure 4.10 shows the pentad-mean annual cycle of OLR, q and ω at the 500-hPa level in four regions of the AMS. For the North American Monsoon the seasonal cycle of OLR, q and ω is relatively well represented in the simulations. During late boreal winter and early spring, OLR increases steadily as a result of surface warming. However, in early June, near the onset date (Douglas et al., 1993; Geil et al., 2013), OLR sharply decreases reaching a minimum value of 246 W m^{-2} by mid-July. The vertical velocity decreases steadily from January to a minimum in August, indicating ascent from May 1st until September 15th. The models show similar seasonal cycles but overestimate the summertime OLR by $\approx 6 \text{ W m}^{-2}$ and underestimate mid-level moisture by 0.3 g/kg and ω by 0.01 Pa s^{-1} . The simulated shallower convection and drier mid-troposphere is seemingly compensated by stronger ascent.

In the MSD region, OLR and q show signs of convective activity from mid-April, as OLR sharply decreases and moisture increases. The characteristic MSD bimodal distribution of precipitation can also be observed as two peaks of low OLR, high q and low ω . These periods are separated by a period of relatively higher OLR, lower q and weaker ascent from June 15 until late August. Although arguably with a small

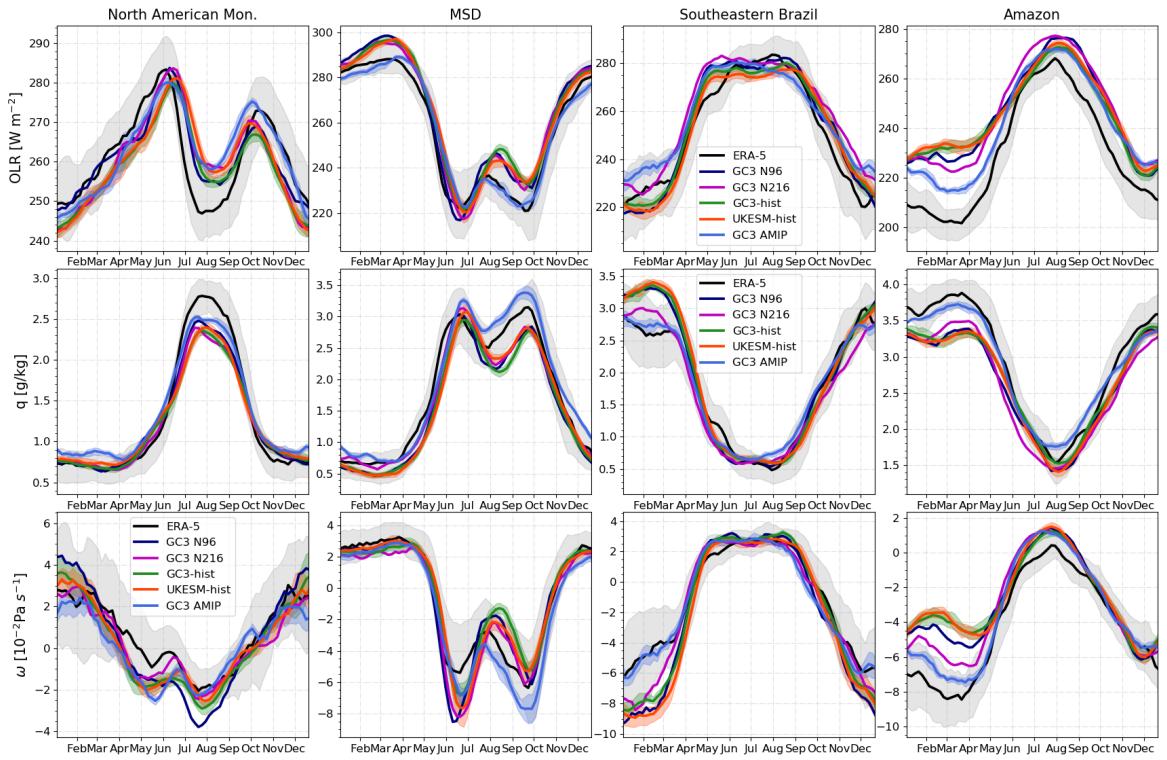


Figure 4.10: Pentad-mean (upper) out-going longwave radiation (OLR), (middle) specific humidity at 500-hPa and (lower) ω 500-hPa. These are shown from left to right for the North American Monsoon, the Midsummer drought, southeastern Brazil and the core Amazon. The uncertainty in ERA-5 data, shown as faint gray shading was estimating by bootstrapping with replacement the ERA-5 record 10,000 times.

dry bias with shallower convection after mid-July, the simulations follow closely the observed seasonal cycle.

The simulated conditions during the first peak period show similar OLR and mid-level moisture but stronger ascending motions, which may explain the positive rainfall bias in this period showed in Figure 4.9a. In the period between the first peak and the MSD, the simulated OLR increases more sharply than observations from 220 W m^{-2} (June 15) to 250 W m^{-2} (early August), with similar behaviour in ω and q , which may also be related to the strong MSD precipitation differences described in the previous section. The period during the second peak of rainfall in September shows signs of shallower convection and a drier mid-level when compared to ERA5.

In southeastern Brazil, the simulations reasonably follow the timings of the annual cycle of OLR, q and ω of the reanalysis, particularly during austral winter. The moisture q in ERA5 during the dry seasons of austral fall, winter and spring is reasonably simulated by all the experiments. However, during austral summer, the coupled model simulations show significant biases characterised by stronger ascent

and increased specific humidity in the mid-levels, although the height of convection (OLR 225 W m^{-2}) is only modestly higher in the simulations.

The simulated OLR, q and ω exhibit the highest biases in the Amazon. During austral summer, particularly January and February, the simulated convective activity is shallower (OLR bias of $+25 \text{ W m}^{-2}$) and weaker (positive ω bias $+0.02 \text{ Pa s}^{-1}$) and the mid-level troposphere is drier (-0.5 g/kg) than in ERA5. All these biases are in agreement with the dry Amazon bias described in the previous section. In spite of biases in the magnitude of OLR, q and ω during peak convective activity, the seasonal variation is very well simulated so that convective activity, as evidenced by these metrics, starts and ends in the simulations within one or two pentads of the reanalysis. The smallest biases in coupled simulations are those of GC3 N216-pi, not just for the Amazon region but for the other regions as well. The simulated OLR, q and ω by GC3-amip in southeastern Brazil and the Amazon show a much better agreement with the reanalysis during austral summer than the rest of the observations.

4.5 ENSO Teleconnections

El Niño-Southern Oscillation (ENSO) teleconnections are the prominent source of interannual variability for the AMS (Vera et al., 2006), as summarized in section 2.4. The response to ENSO events is investigated in this section, which first shows the temperature, sea-level pressure (SLP) and precipitation responses to observed and simulated ENSO events in the AMS, to then analyse the effect of ENSO flavours on the AMS. Finally, results show a possible influence of the QBO for the teleconnections of ENSO.

Throughout this section, ENSO events were defined when the DJF-mean Niño 3.4 index was above or below 0.65 (Trenberth, 1997). Other indices, including the use of a 5-month running mean (Trenberth et al., 1998), were tested without significantly changing the results. Previous studies (e.g. Menary et al., 2018; Kuhlbrodt et al., 2018) showed that the MOHC models reasonably simulate several characteristics of ENSO such as the period and SST patterns.

4.5.1 Canonical teleconnections

The surface temperature and sea-level pressure (SLP) responses to ENSO events are shown in Figure 4.11 for HadGEM3, UKESM1 and ERA5 data during DJF, the season of strongest impact of ENSO events. The characteristic warm anomaly during

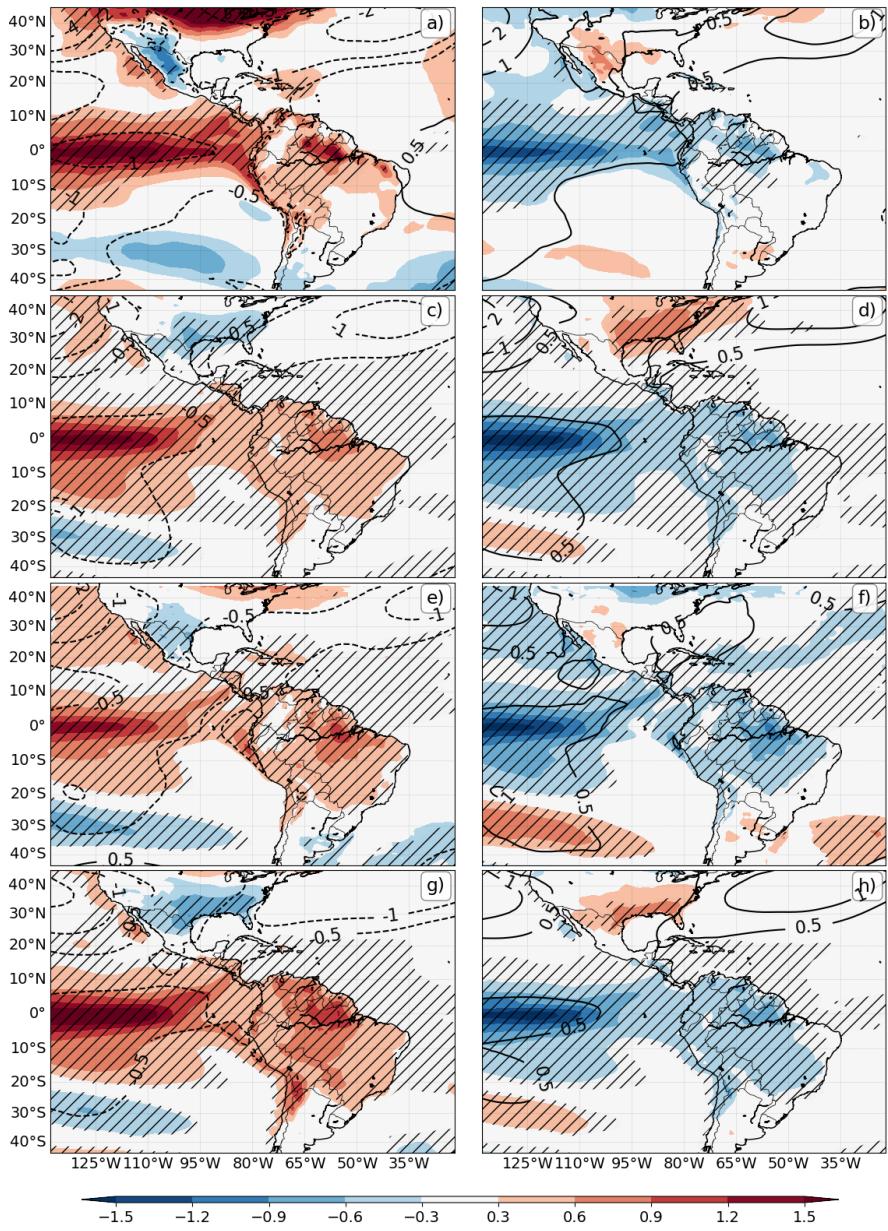


Figure 4.11: DJF Temperature anomalies (colour contours in K) and SLP (line contours in hPa) during (a, c, e, g) El Niño and (b, d, f, h) La Niña events. Results are shown for (a, b) ERA-5, (c, d) UKESM1-histl, (e, f) GC3 N96-pi and (g, h) GC3 N216-pi. The hatched regions denote 99% significance from a Welch t-test for the temperature field.

El Niño events in the East Pacific Ocean does not extend to the east in all the simulations as the observed warm anomaly. In turn, the cold anomalies during La Niña events in the Central Pacific are colder in the simulations than in ERA5. The teleconnection to southern North America, i.e., colder (warmer) conditions in southern

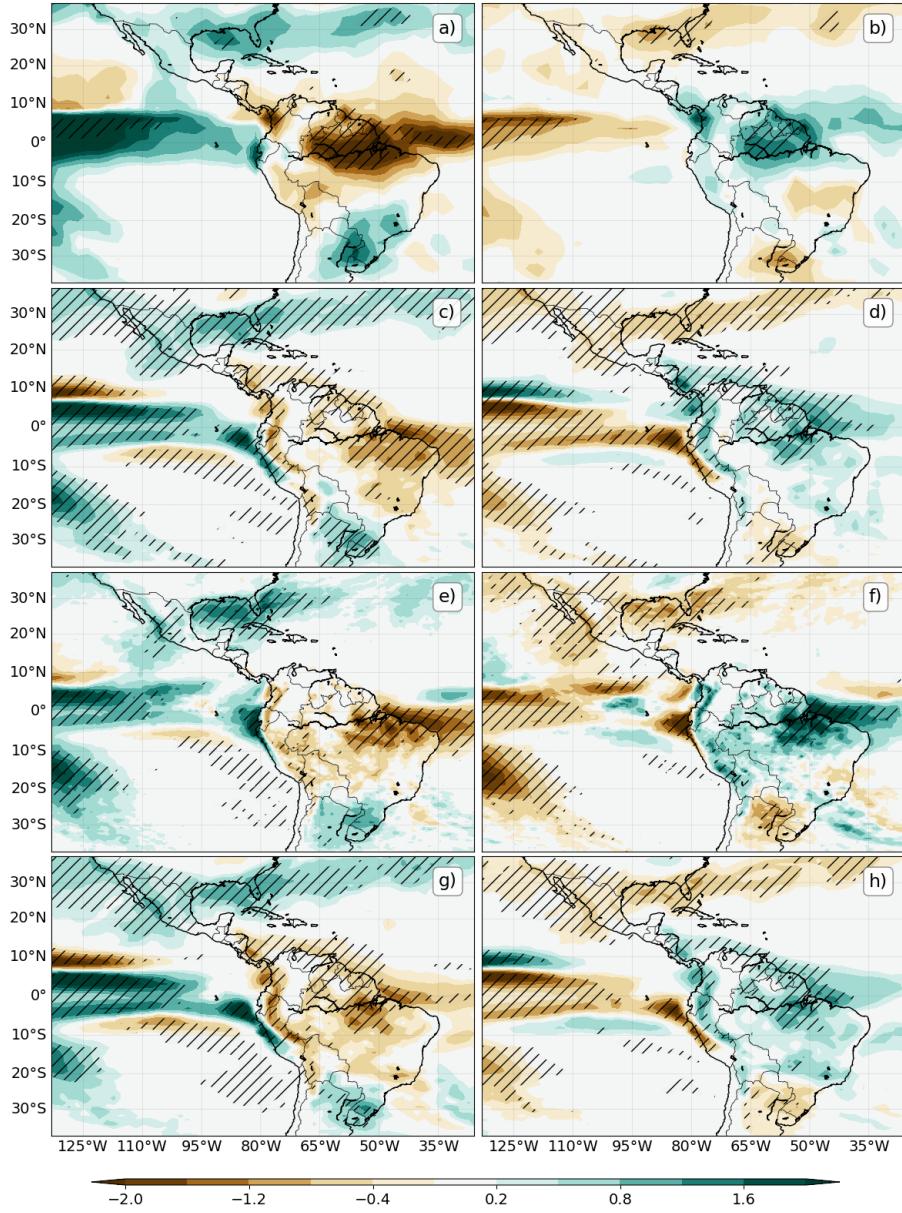


Figure 4.12: As in Figure 4.11 but for the rainfall response [mm day^{-1}] using GPCP as the observational dataset.

(northern) North America during El Niño events are relatively well simulated. For example, the simulated and observed teleconnection patterns to South America, e.g., the cold anomalies during La Niña events in northern South America are well simulated. However, the low resolution simulations show a broader and stronger than observed negative response in southeastern US to El Niño events.

The SLP response in the north Pacific and North America, known as the Pacific North-American pattern, is linked with a displacement of the subtropical jet affect-

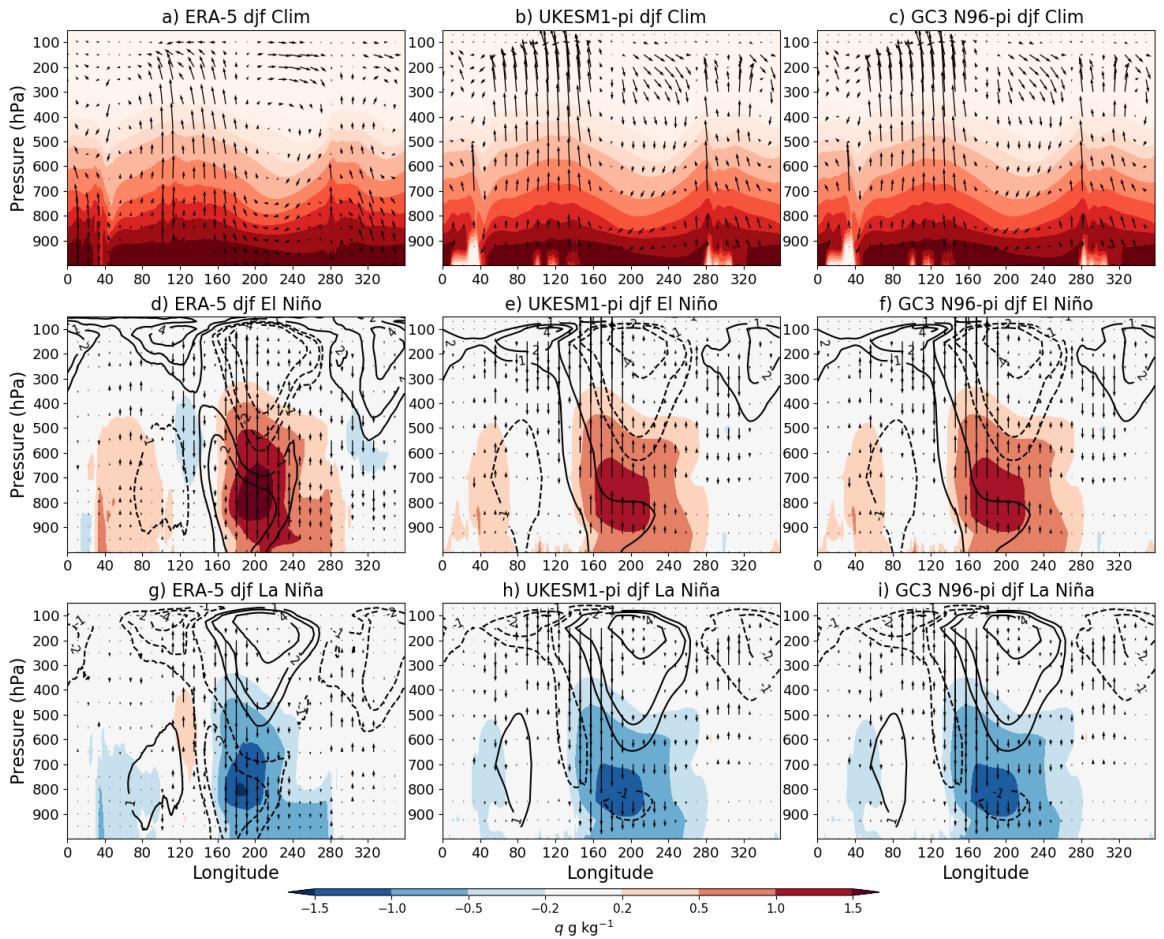


Figure 4.13: DJF Longitude-height Walker circulation anomalies of specific humidity (colour-contours), ω (vectors) and zonal wind (line-contours) during El Niño events (left) and La Niña events (right). Results are shown for ERA-5 (upper), UKESM1-pi (middle) and HadGEM3 piControl (lower).

ing the eastward propagation of wave activity that reaches the North Atlantic (e.g. Bayr et al., 2019; Jiménez-Esteve and Domeisen, 2020). During El Niño events, the Aleutian Low is strengthened in ERA5, with a strong SLP anomaly (-4 hPa) off the coast of California. The models show a similar but smaller SLP response in the same region. El Niño events are associated with a negative phase of the North Atlantic Oscillation (NAO), with an opposite response for La Niña events. While the models seem to be able to capture this response of the NAO, the simulated response is weaker than observed. A sensible representation of the ENSO-NAO tropospheric teleconnection may be relevant to then simulate the effect of the NAO on Central American and northern South American rainfall (Giannini et al., 2000, 2004).

The rainfall anomalies associated with ENSO events are shown in Figure 4.12.

Three regions in the AMS have a significant precipitation response to ENSO events in the observations and simulations. In southern North America, rainfall increases (decreases) during El Niño (La Niña) events due to the effects of the PNA pattern on the subtropical jet, which influences the frequency and latitude of propagation of wintertime midlatitude disturbances (Vera et al., 2006; Bayr et al., 2019). The GPCP dataset (Figure 4.12a, b) shows significant boreal winter rainfall increases in southeastern US and the Gulf of Mexico during El Niño events, and an opposite response to La Niña phases. All the simulations reproduce this teleconnection rainfall pattern. The models also simulate the observed response in southeastern South America (SESA) of positive anomalies during El Niño and negative anomalies during La Niña events. This teleconnections is also associated with the effect of ENSO on midlatitude and subtropical jet activity, but for the Southern Hemisphere.

The anomalies in the Amazon show the strongest response to ENSO events in the observations. Significant positive (negative) rainfall anomalies during the negative (positive) phase of ENSO in northern South America are observed in GPCP. All the simulations show a very similar and statistically significant response. This teleconnection works through the coupling of ENSO with the Walker circulation (Vera et al., 2006; Cai et al., 2019), which is illustrated in Figure 4.13.

The climatological Walker circulation during DJF shows strong ascent in the 100-160°E and the 280-310°E regions, which correspond to the maritime continent and South America (Figure 4.13a). During El Niño events, there is increased specific humidity throughout the lower troposphere in the Central and Eastern Pacific, associated with ascending motions in this region and negative low-level wind anomalies and positive upper-level wind anomalies (Figure 4.13d). In other words, an eastward shift of the Walker circulation. The wind, vertical velocity and specific humidity anomalies are the opposite during La Niña events, indicative of a stronger Walker circulation, slightly shifted to the west. The models seem to broadly reproduce the observed changes to the Walker circulation during ENSO events (Figure 4.13).

Figure 4.14 shows the observed and simulated precipitation responses in four regions of the AMS to different magnitudes of ENSO events, by binning events for their magnitude of the EN3.4 index and the corresponding precipitation anomaly from the climatology in each region. This figure aims to show the degree of linearity of ENSO teleconnections to the AMS. While the observed response shows some degree of linearity for El Niño events in South America (panels c, d), the majority of the observed responses, particularly to La Niña phases, are not linear.

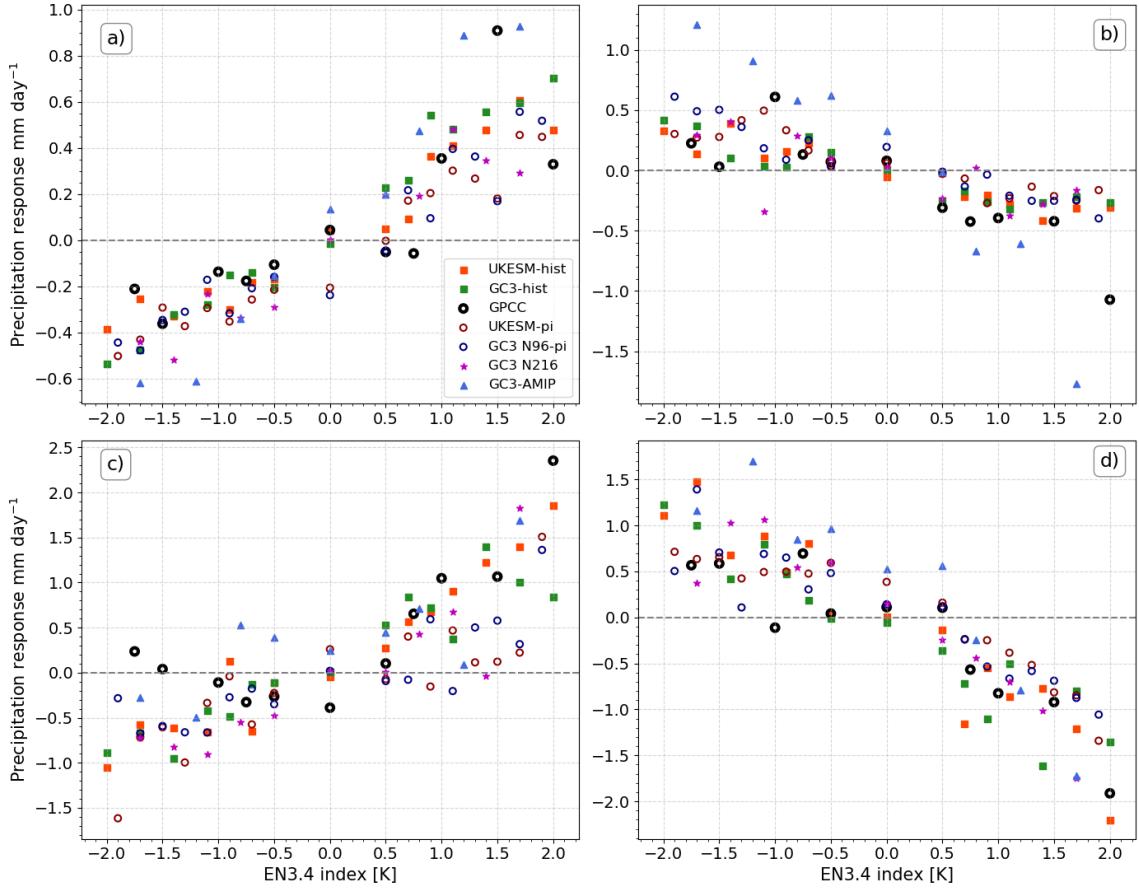


Figure 4.14: Precipitation response [mm day^{-1}] as a function of the El Niño 3.4 index (see text) for (a) southwestern North America [$20\text{-}37^\circ\text{N}, 112\text{-}98^\circ\text{W}$], (b) Central America and southern Mexico [$5\text{-}19^\circ\text{N}, 95\text{-}83^\circ\text{W}$], (c) South Eastern South America [$35\text{-}25^\circ\text{S}, 60\text{-}50^\circ\text{W}$], and (d) the Amazon [$10\text{-}0^\circ\text{S}, 70\text{-}45^\circ\text{W}$]. The observation scatter points are from GPCC in the period of 1940–2013.

However, the simulations show several signs of linearity. For instance, consider the historical experiments, UKESM1-hist and GC3-hist, which show that the precipitation responses in southwestern North America, SESA and the Amazon increases roughly linearly as the magnitude of SST anomaly increases. In contrast, some other simulated responses, e.g. to La Niña phases in South America in the piControl simulations, show signs of non-linearity.

4.5.2 The role of ENSO flavours

As described in section 2.4, not all ENSO events are observed with the same SST anomaly pattern in the Pacific Ocean. These different SST patterns for each ENSO event are considered to be a source of non-linearity of ENSO impacts over South

America (Sulca et al., 2018; Cai et al., 2020). Principal component analysis has shown that ENSO events may be separated into two categories: Central Pacific (CP) and East Pacific (EP) events (Cai et al., 2020), which highlight where the peak SST anomaly is found in the Pacific Ocean. Figure 4.15 shows that both UKESM1 and GC3 reasonably simulate the observed SST patterns associated with EP and CP El Niño events, although the simulations show CP SST patterns to spread further to the east than the HadSST dataset. The simulations are also able to replicate very broadly the observed differences in the frequency of each event as CP La Niña events are more frequent than EP La Niña events, while the opposite is true for El Niño events.

Furthermore, Figure 4.16 compares the precipitation anomalies for each type of

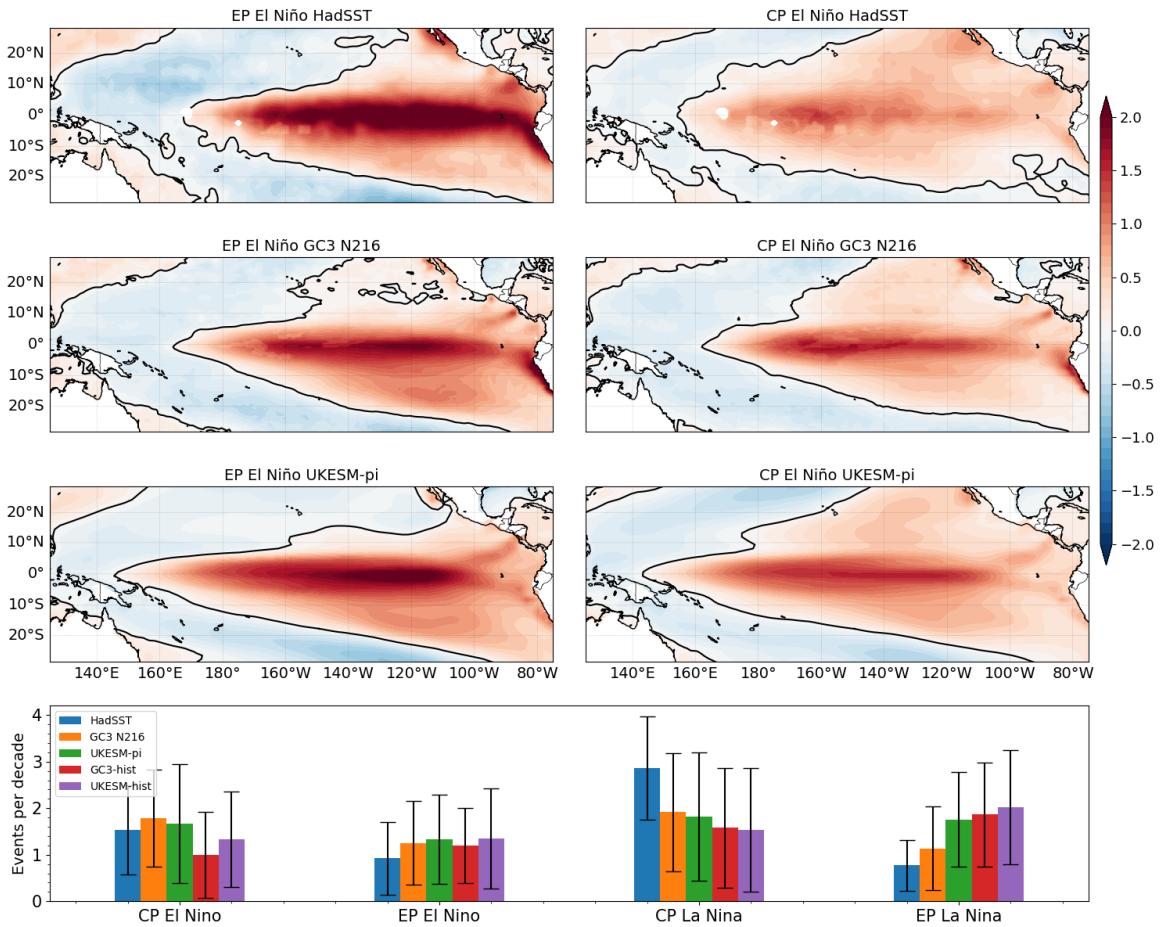


Figure 4.15: SST anomalies [K] for East Pacific (EP) and Central Pacific El Niño events in HadSST, GC3 N216 and UKESM piControl. EP (CP) events were defined where the E-index (C-index) was greater than 1. In the bottom panel, the frequency of events per decade (with standard deviation as error bar) is shown for HadSST and the simulations used in this study. The E-index is computed from $(PC1 - PC2)/\sqrt{2}$ and the C-index from $(PC1 + PC2)/\sqrt{2}$.

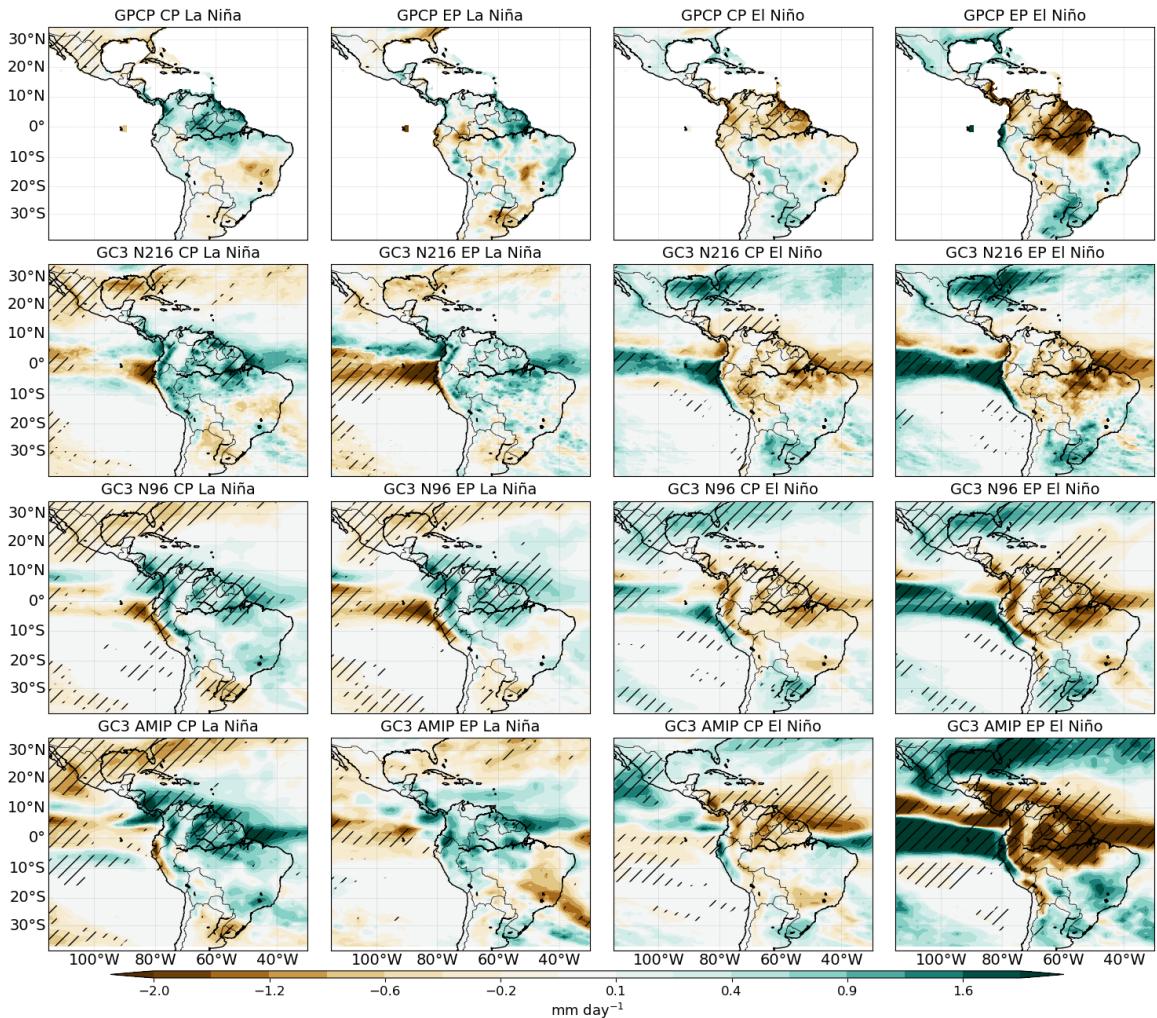


Figure 4.16: Precipitation anomalies in GPCC 1940-2013, GC3 N216-pi, GC3 N96-pi and GC3 AMIP for the four different types of ENSO events, as defined by Cai et al. (2020). Statistically significant anomalies (95% confidence level) are hatched.

ENSO event in observations with three simulations: GC3 N96-pi, GC3 N216-pi and GC3-amip. The observed precipitation response in the GPCC dataset to EP La Niña over equatorial South America is not significant and is smaller than the strong positive response to CP La Niña events in the same region. However, the simulated response in GC3 N96-pi and GC3 N216 during La Niña events appears to be more independent of the type of event. In contrast, GC3-amip shows different magnitudes of responses to different types of La Niña events, in particular a positive, and significant, anomaly for CP La Niña events in the Amazon and weaker and not significant anomalies during EP events, which agrees with observations.

The observed response to El Niño events in GPCC is also dependent on the type of

event. EP El Nio events show significant negative anomalies over the Amazon and positive anomalies over SESA whereas CP events only show significant anomalies (-1 mm day^{-1}) over northeastern South America. While the coupled models (GC3 N96-pi and GC3 N216) do show a stronger response to EP El Nio events than to CP events, the patterns of the response are very similar. In contrast, the response in GC3-amip agrees with observations. For this experiment, stronger negative responses to EP El Nio events are observed in the Amazon but the response to CP events is much weaker and is only significant in northeastern South America. In other words, GC3-amip agrees well with the observed non-linear teleconnection patterns whereas the teleconnections in the coupled models do not depend on the type of ENSO event.

4.5.3 A possible influence of the QBO on tropical ENSO teleconnections

Section 2.5 discussed the observational and modelling evidence of the effects on deep convection associated with the stratospheric quasi-biennial oscillation (QBO). In particular, some evidence suggest that the QBO may play a role to determine interannual variability of the Walker circulation and monsoons (Giorgetta et al., 1999; Collimore et al., 2003; Liess and Geller, 2012).

This section evaluates whether the simulations analysed in this chapter, as well as observations, show signs of an influence of the QBO on the AMS. In particular, the analysis aims to understand whether the QBO may be a source of non-linearity and non-asymmetry for the teleconnections of ENSO associated with deep convection and the Walker circulation. In all cases, the phases of the QBO were defined using a 70 hPa zonal mean zonal wind index, with a threshold of $+2 \text{ m s}^{-1}$ for the westerly phase (QBOw) and -2 m s^{-1} for the easterly phase (QBOe).

Composites of the precipitation response to La Niña (LN) events in Figure 4.17 show that the phase of the QBO may be determine the strength and location of the teleconnection. While the precipitation difference in the western Pacific is relatively similar during QBOe than during QBOw in observations and simulations, the teleconnections to Australia, South America and the maritime continent are notably different depending on the QBO phase. In the GPCP dataset, the composite difference QBOe-QBOw during LN events suggests that the characteristic positive precipitation response during LN events in the Amazon, is largely associated with QBOw phases, whereas LN events during QBOe appear to have little effect over South America. A similar result is obtained for GC3 N96-pi.

These precipitation responses are further investigated by changes in the overturning circulation (Figure 4.18). As depicted in Figure 4.13, La Nia events are associated with a westward shift in the Walker circulation with a strengthening of the low-level

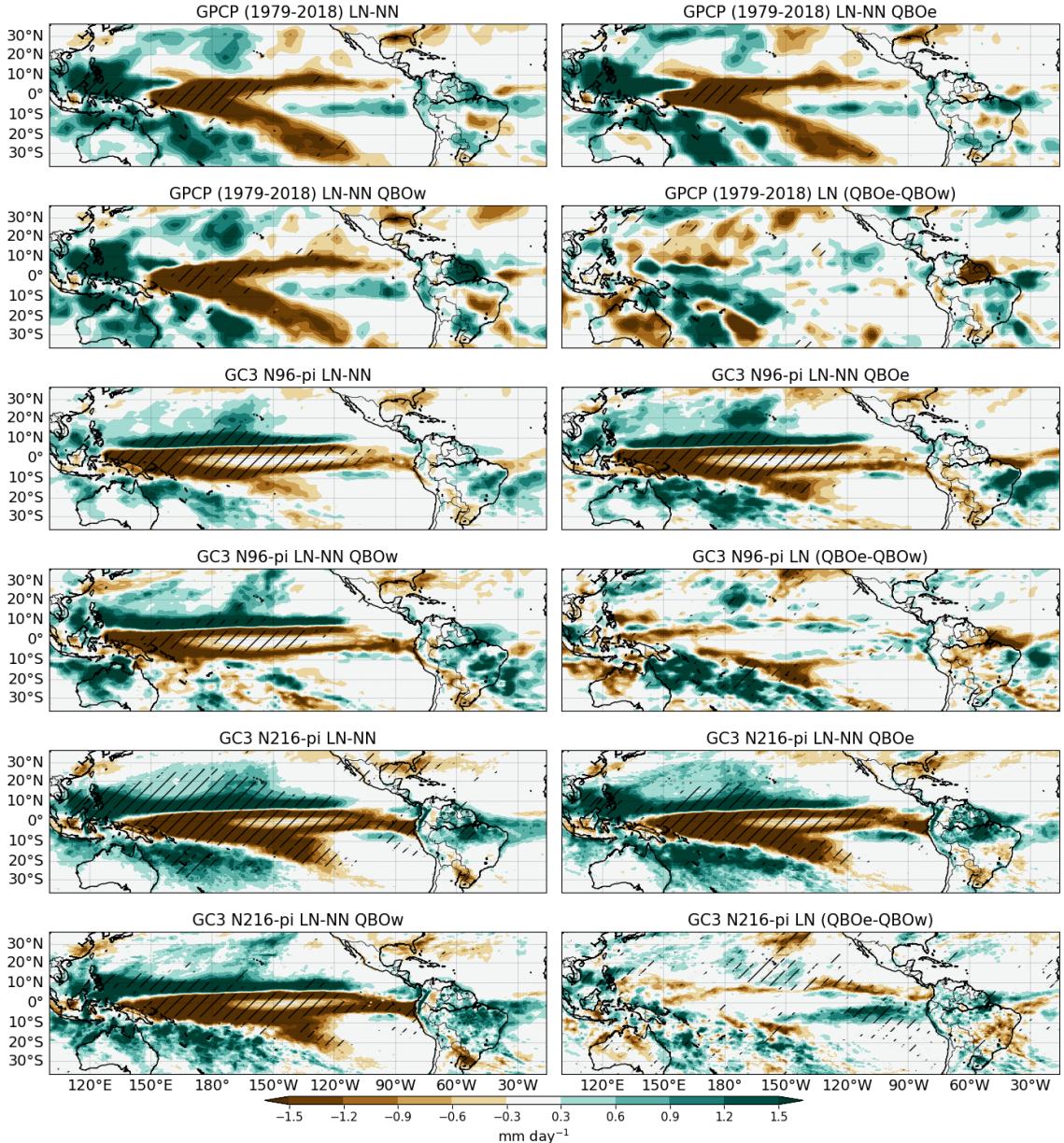


Figure 4.17: Composite precipitation differences during JFMA in GPCP (1979-2018), GC3 N216-pi and GC3 N96-pi between (top) La Nia and Neutral ENSO conditions. The two middle panels show a subset of the top panel, by separating the La Nia composite based on the phase of the QBO. The lower panel shows the differences QBO E-W during La Nia periods. Statistically significant anomalies (95% confidence level) are hatched.

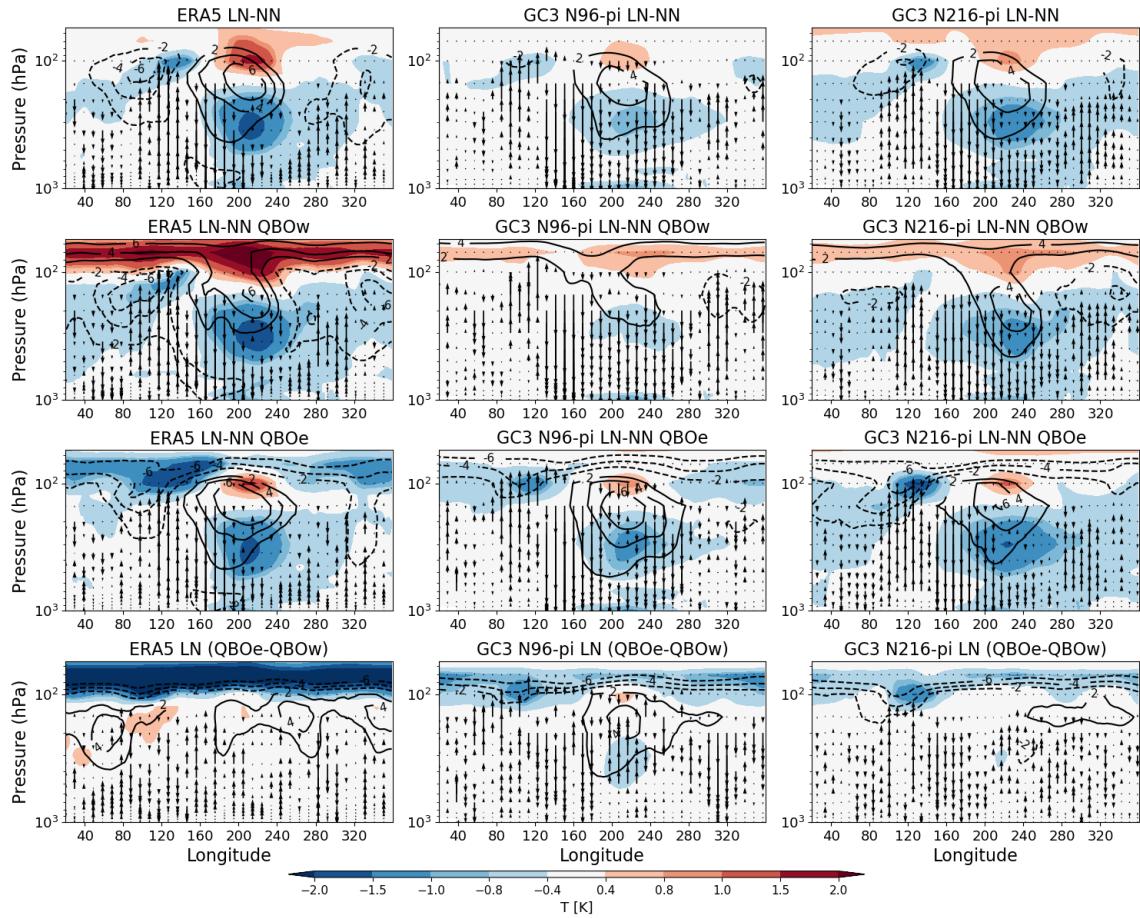


Figure 4.18: Longitude-height differences (JFMA) of equatorial (10S-10N) air temperature (color shading), zonal wind (contours) and vertical velocity (ω - vectors). The differences shown from top to bottom are between all La Niña (LN) periods and Neutral conditions (NN), between LN and NN during QBOw, LN-NN during QBOe, and the difference between LN events on different QBO phases (LN QBOe-QBOW).

easterlies in the Pacific Ocean. Figure 4.18 shows that during LN the tropical troposphere cools and the UTLS region in the Central Pacific warms. These temperature anomalies are weaker in the simulations than in ERA5.

The zonal wind anomalies in the upper-troposphere associated with LN events show different patterns and strengths during QBOw than during QBOe. The mean teleconnections during LN show positive upper-tropospheric anomalies above the Pacific Ocean, but these anomalies are stronger during QBOe than during QBOw in ERA5 and the two simulations shown. In ERA5, most of the upper troposphere shows positive zonal wind differences in the QBOe-QBOW panel.

There are three regions where ascending and descending motions are more greatly affected by LN events: the maritime continent, the Pacific Ocean and South America.

The observed effect of the mean LN teleconnection is the following: anomalous ascent is seen in the maritime continent and in South America, in agreement with a stronger Walker circulation, whereas anomalous descending motions is observed in the Central and eastern Pacific associated with a westward shift of the Walker circulation.

The effect of LN over ascending and descending motions is seemingly also affected by the QBO phase, according to the bottom panels of Figure 4.18. In ERA5 and the simulations, the anomalous ascent observed in South America during LN events is mostly associated with QBOw, whereas only small anomalous ascent is observed during QBOe. However, ERA5 disagrees with the simulations in the western Pacific region (140-180E), as the simulations suggest larger anomalous descent during QBOe than during QBOw, whereas in ERA5 these descending anomalies are larger during QBOw.

A similar analysis was conducted to evaluate the effect of the QBO during the positive and the neutral phases of ENSO. These results are not shown because, although tentative suggestions were found that the QBO may play a role during these other phases of ENSO, there was little agreement between the models and ERA5/observations. Furthermore, the QBO representation in these CMIP6 models is biased in the UTLS region. In particular, the temperature signal associated with circulation of the QBO, most clearly seen in the bottom panels of Figure 4.18, is much weaker in the models.

As suggested by the literature summarised in section 2.5, this temperature signal could be the key aspect of any effect of the QBO on deep convective systems, and as such, the evidence from a short record (ERA5) or models with key biases in possible processes involved presented in this chapter warrants both caution and more work. This topic will be investigated in the next chapters.

4.6 Summary and discussion

This chapter analysed the output from the UK MOHC models, HadGEM3 and UKESM1 in their pre-industrial control, historical and AMIP experiment contributions to CMIP6. In particular, the chapter evaluates several aspects the simulated climate related to the AMS. The use of the output of UKESM1 and two resolutions of HadGEM3 allowed the direct comparison of the effect of including Earth System processes or increasing resolution for representing regional monsoon rainfall. A schematic in Figure 4.19 shows the primary components of the AMS climate and summarizes the main biases found in these simulations and this chapter.

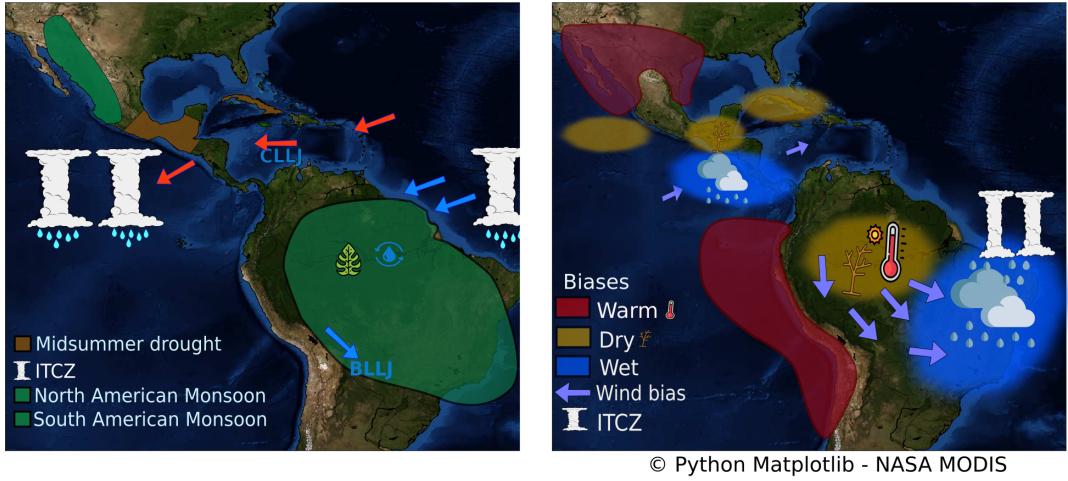


Figure 4.19: Schematics of (a) the main features in the AMS and (b) the main biases in UKESM1 and HadGEM3. In (a) the boreal summer easterlies (red) and austral summer circulation (blue) are shown with the Caribbean and Bolivian Low-level Jets (CLLJ and BLLJ, respectively). In (b) the biases are shown for the respective northern and southern Hemisphere summers. The ITCZ bias in (b) refers to the southward displacement bias of the Atlantic ITCZ in the simulations.

Rainfall in the North American Monsoon was particularly well simulated by the models. The seasonal cycle, peak monsoon rainfall rates and timings of monsoon onset and retreat in the simulations agreed well with TRMM. The historical experiments overestimate the mean temperature in most of the Americas by 1.5 K, but particularly in boreal summer in southwestern North America (+4 K). In spite of this warm bias, the temperature seasonal cycle is well represented by these models.

These results suggest model improvement on the simulation of the North American Monsoon from previous versions of the MOHC models (Arritt et al., 2000), and most of the model cohorts of CMIP3 and CMIP5 (Geil et al., 2013). For example, most of CMIP5 models showed a very wet bias during monsoon maturity whereas rainfall during monsoon maturity in all the experiments of this chapter are within less than 1 mm day⁻¹ of observations, during the maturity stage. However, these models continue to show biases during monsoon retreat as rainfall does not decrease as sharply as in observations after mid-September, which suggests a continued bias in the winter-time precipitation associated with cold-fronts (Adams and Comrie, 1997).

The Midsummer Drought (MSD) of southern Mexico and Central America is a regional feature of precipitation that most of CMIP5 models had difficulty capturing, with the MOHC models being amongst the few exceptions (Ryu and Hayhoe, 2014). The MSD in UKESM1 and GC3 continues to be relatively well represented, which is

a remarkable feature of the models compared to others. However, the experiments analysed in this chapter showed various differences in the timing and strength of the bimodal cycle when compared to observed gridded-datasets and ERA5.

The models simulate a wetter-than-observed first peak of precipitation and a drier MSD period, therefore simulating a larger difference between the first peak and the dry period. While in observations this difference between the first peak and the MSD period ranges between 2-3 mm day⁻¹, in the simulations this difference is closer to 6 mm day⁻¹. Rainfall during the first peak has been too wet in these models since CMIP3, suggesting a persistent wet bias in this region, likely associated with the bias in East Pacific ITCZ also shown in this chapter and in recent studies (Ryu and Hayhoe, 2014; Mulcahy et al., 2018). In contrast, the so-called second peak of precipitation, observed in late August is simulated in close agreement with TRMM, except in the AMIP experiment, which has a wet bias of 2 mm day⁻¹ at this stage.

The capability of these models to simulate a bimodal regime of precipitation raises possibilities to further use these simulations to understand the mechanisms as to why the MOHC do represent a MSD regime but other models fail. Furthermore, section 2.3 discusses several open questions regarding the mechanisms that cause the observed MSD. Therefore, the results of this chapter suggest that these models may be further explored to investigate the processes that explain the timing, strength and overall existence of the MSD precipitation regime.

The East Pacific ITCZ migration and position was shown to be relatively well represented by the models (Figs. 4.4 and 4.5). However, the models showed an overestimation of boreal summer rainfall near the coast of Central America (Figure 4.9). These biases are associated with an easterly bias in the low-level wind, suggesting a bias in the flow from the Caribbean Sea into the Eastern Pacific (Herrera et al., 2015; Durán-Quesada et al., 2017). The simulations also showed a biased Atlantic ITCZ that was displaced south of the observed ITCZ position during boreal winter (Figure 4.5), particularly in the low resolution coupled simulations.

In the Amazon, the simulations showed a warm bias (+2 K) during austral spring and summer, a typical feature of previous models (Jones and Carvalho, 2013), and a colder than observed southeastern Brazil. These biases were linked with decreased cloud cover and less rainfall over the Amazon and more high clouds and rainfall in southeastern Brazil (Figures 4.8 and 4.10). The low cloud cover, warm and dry Amazon biases are intertwined with the low-level circulation from the Atlantic into the South American continent. The biases in the circulation during austral summer were observed as a northerly flow anomaly over the central and southern Amazon, a

feature that has been associated with a stronger moisture transport away from the Amazon (Marengo et al., 2012; Jones and Carvalho, 2018).

During the period of maximum rainfall rates in February, the simulations can overestimate rainfall by 3 mm day^{-1} in southeastern Brazil and underestimate rainfall in the Amazon by a similar rate. The historical experiments showed a small drying response to historical forcing in the Amazon therefore slightly increasing the magnitude of this dry bias compared to the unforced piControl experiments. The AMIP simulation with the SST biases removed improved the representation of the Atlantic ITCZ and the precipitation, cloud cover and temperature biases over the South American Monsoon. The improvement in the circulation and precipitation biases in the AMIP simulation suggest that the origin of the dry Amazon bias are the biases in the Atlantic SSTs.

The canonical teleconnection responses of temperature, SLP and precipitation in the AMS to ENSO events are well represented in these models. For example, the simulated spatial patterns and strength of the positive (negative) precipitation anomalies observed in northern Mexico and South Eastern South America during El Niño (La Niña) agree well with observations and reanalysis. Similarly, the teleconnection to the Amazon is well represented for both phases of ENSO, in spite of relevant biases in the mean state of the South American monsoon discussed above.

ENSO teleconnections in these simulations were found to be approximately linear, i.e., the precipitation response is linearly related to the magnitude of the SST perturbation in the EN 3.4 region. These experiments also show signs of symmetric teleconnections as positive and negative phases produce the opposite and equivalent precipitation response in the AMS. In contrast to observations and the GC3 AMIP simulation, the precipitation response in the coupled models appears to be independent of the type or flavour of ENSO events into Central and East Pacific events. The fact that these models show a reasonable representation of ENSO diversity in SST patterns but the models do not replicate the observed non-linear dependance to ENSO events warrants further analysis.

The main biases, depicted in 4.19, in these experiments are generally smaller in the medium resolution GC3 N216 compared to the low resolution experiments, which suggests improved model performance with increased horizontal resolution. In contrast, including Earth System processes in the UM model only affects the surface temperature response to historical forcing and not the dynamical biases that drive the precipitation and ITCZ biases.

In short, the main dynamical biases in UKESM1 are very similar to those in GC3-N96 as these two models share the same dynamical core; only when resolution is increased these biases are reduced notably. A noteworthy difference between UKESM1 and GC3 is that warming over the historical period in Mexico and the Amazon is higher in UKESM1 than in GC3. In general, UKESM1-hist shows a stronger temperature response to forcing than GC3-hist, as UKESM1 has been reported to have a greater climate sensitivity than GC3 (Andrews et al., 2019; Sellar et al., 2019). This differential warming may be a consequence of the land-use change in these regions playing a role in the UKESM1 representation of soil-atmosphere feedbacks.

The improvement in the medium resolution simulation compared to the low-resolution simulations may be associated to the improved dynamics of the ocean or the atmosphere. For example, the Atlantic ITCZ biases have been shown to be directly affected by processes in the convective scheme (Bellucci et al., 2010), such as the treatment of entrainment and moisture-cloud feedbacks (Oueslati and Bellon, 2013; Li and Xie, 2014). The resolution of the ocean model has been shown to impact the eddy heat flux parametrisation and the associated heat uptake and transport of the ocean (Kuhlbrodt et al., 2018). The improvement in the Atlantic SSTs and ITCZ and the associated dynamics in GC3 N216-pi also improves the associated circulation biases and moisture transport in the South American Monsoon. In other words, the oceanic resolution may play an important role in the cross-equatorial heat and moisture transport, SST gradients over the equatorial Atlantic which in turns improves the land-sea circulation over the Amazon during austral summer, a circulation that is key for representing the spatial distribution of rainfall in South America.

Chapter 5

On the physical mechanisms of the MSD of southern Mexico and Central America in the Met Office Unified Model

The physical mechanisms responsible for this bimodal signal have remained elusive in spite of active research and debate still exists over the dominant mechanisms that explain the seasonal cycle of regional rainfall patterns.

This chapter revisits the problem of understanding the physical mechanisms that give rise to the Midsummer drought by comparing reanalysis and observational datasets with a number of experiments from the CMIP6 experiments for the MOHC UM model under different configurations. The roles of the East Pacific sea-surface temperatures and surface fluxes, the Caribbean Low-Level Jet and cloud-radiative effects are investigated. A thermodynamic diagnosis of the MSD is also performed via the analysis of the moist static energy budget and the surface entropy on both reanalysis and model data.

5.1 Introduction

A double-peak signal in the climatological seasonal cycle of precipitation has been documented in several regions of southern Mexico, Central America and the Caribbean (Mosíño and García, 1966; Magaña et al., 1999; Gamble et al., 2008) but an accurate depiction of the physical mechanisms that give rise to this bimodal signal remains elusive. Studies disagree on the the roles of local cloud-radiative feedbacks, thermodynamic and dynamic mechanisms associated with regional climate features.

Section 2.3 summarizes the literature that has aimed to explain the physical mechanisms that increase precipitation at the start and end of summers but decrease during the midsummer, thus called Midsummer drought (Magaña et al., 1999). The vast majority of studies are observational in nature and analyzed monthly-mean changes to the precipitation. Little-to-no attention has been given to this problem in global coupled climate models because most of them do not simulate the basic features of the seasonal cycle in this region.

Ryu and Hayhoe (2014) shows that out of all the models in the CMIP3 and CMIP5 cohort only a handful show signs of a bimodal signal, with the MOHC (HadGEM2) model being amongst the few that could simulate the MSD. Chapters 4 and 5 show that the CMIP6 MOHC simulations reproduce the timings and strength of the bimodal signal of precipitation with reasonable skill, albeit with a stronger first peak and a later onset of the MSD (MSDO). The characterisation of the mechanisms that produce the bimodal signal in model data is therefore another avenue to understand the real-world MSD, and one that has not been explored before to much detail. Furthermore, characterising the differences between experiments with varying forcing and different horizontal resolution will also provide relevant information about the simulated current and future response to greenhouse forcing as well as highlight relevant processes that give rise to biases in the region for model development.

The previous chapter provides a useful diagnostic tool to separate the different stages of the MSD, a wavelet transform method (WT) that diagnoses the onset and end of the rainy season, as well as the start and end of the MSD. Similarly, the WT method can provide a robust measure of whether a given year and/or grid-point exhibits a robust bimodal signal. This chapter will use the WT method to separate the different stages of the MSD in model and observational data.

Most conceptual theories for the MSD involve elements of the regional climate of the Intra-Americas Sea (IAS) and the East Pacific Ocean, particularly, the Caribbean Low-Level Jet (CLLJ) and the East Pacific sea-surface temperatures (SSTs) are key elements of most theories (Gamble et al., 2008; Herrera et al., 2015). Previous mechanisms proposed by Magaña et al. (1999) and (Magaña and Caetano, 2005) argue for a cloud-radiative feedback with the East Pacific SSTs playing a large role in modulating precipitation through latent heat fluxes. In contrast, other studies argue for a dynamic cause for the MSD associated with the moisture transport responding to changes in the Caribbean-Low Level Jet. This chapter will therefore explore these aspects of the regional climate first.

Convective quasi-equilibrium (CQE) and moist static energy (MSE) budgets have been successfully implemented in monsoon regions to understand mechanisms that drive precipitation changes. The remaining section of this chapter uses MSE and CQE diagnostics to evaluate the role of thermodynamics and ascent within the moist column of the MSD for the seasonal changes to precipitation.

5.2 Data and methods

5.2.1 CMIP6 data

This chapter uses the output from HadGEM3 GC3.1 at N96 and N216 horizontal resolutions and from UKESM1, described in 3.2.2.

5.2.2 Determination of the timings of the MSD

Chapter 5 describes a wavelet transform (WT) method that can determine the timings of the MSD in observational gridded datasets, reanalysis and climate model precipitation time-series. This chapter will use throughout the WT method to determine the onset (MO) and retreat (MR) of the monsoon rainy season, as well as the onset of the MSD (MSDO) and end (MSDE). Similarly, the timings of the first (1P) and second (2P) peaks of precipitation are determined from the WT method.

As a brief reminder of the method, MO and MR are determined by the maximum and minimum sum of WT coefficients computed from 24 to 54 pentads as dilation scales. After MO and MR are determined, a second WT is computed with dilation scales to 10 to 24 pentads and the minimum sum of the WT coefficients corresponds to the onset of the MSD and the maximum to the end of the MSD (MSDE).

5.3 Climatological features

Previous chapters have shown that the seasonal cycle of precipitation is reasonably well simulated by several experiments of CMIP6 over the broad region of the MSD. Figure 5.2 shows the seasonal cycle for the historical and piControl experiments of GC3 N96, GC3 N216 and UKESM as well as the amip experiment of GC3 in the study region of this chapter which encloses the regions with a large amount grid-points that exhibit a robust bimodal signal. The two-peak structure of the MSD is observed in TRMM and ERA5 as two precipitation maxima occurring first during early to mid-June and the second peak at the end of September.

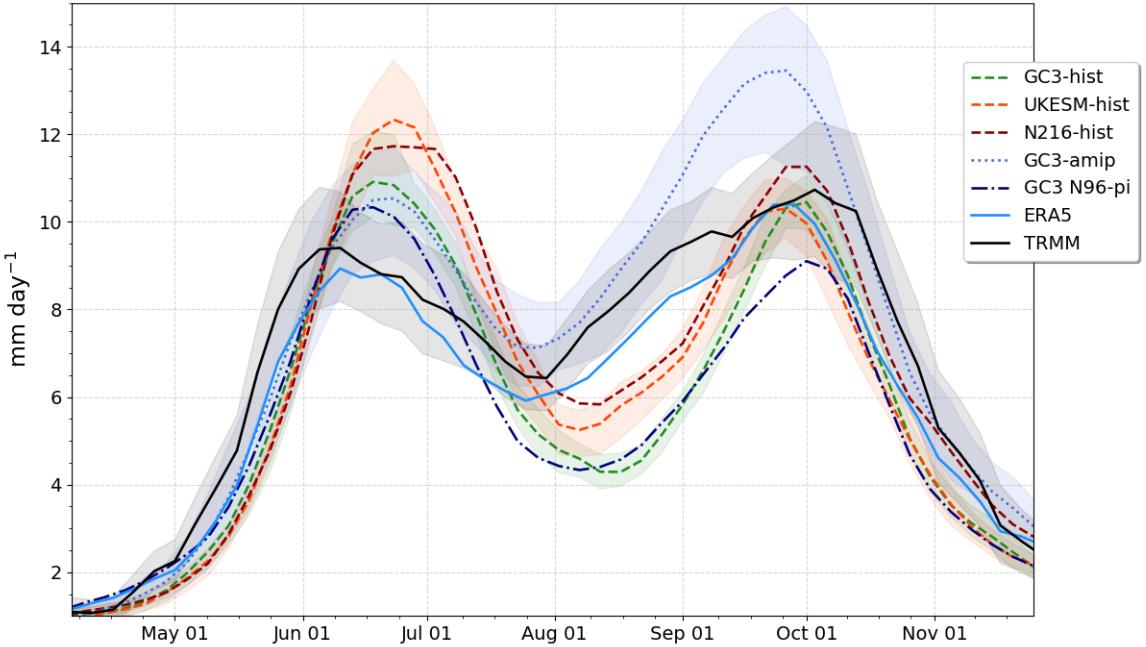


Figure 5.1: Pentad-mean precipitation in southern Mexico and northern Central America [95–86°W, 11–19°N]. Shading shows for TRMM the uncertainty obtained by bootstrapping the interannual variability 10000 times and for the CMIP6 experiments the ensemble spread.

The MSD occurs in the experiments generally one or two pentads after TRMM and ERA5 (see previous chapter) and shows a stronger change of precipitation between the first peak and the Midsummer. In general, the pre-industrial control experiments show a higher magnitude of precipitation during the first peak and the MSD period whereas GC3-amip experiments show a relatively weak magnitude of the first peak but a stronger second peak. This section characterises the main aspects of the MSD and these regional features in the deck experiments of the MOHC.

The seasonal cycle of precipitation in in Central America and Mexico (Fig. 5.2) fits that of a monsoon, i.e., a dry winter and a wet summer season. However, two precipitation maxima, in June and September, are separated by a decrease in precipitation during July and August, *i.e.* the MSD and the difference between the first peak (June 15) to the driest pentad of the MSD (Aug 01) is of about 2 mm day^{-1} , according to TRMM.

Precipitation in these regions depends on several factors such as the seasonal migration of the East Pacific (EP) and Atlantic ITCZs. The SSTs in the Gulf of Mexico, the Caribbean Sea, the western tropical Atlantic and the Eastern Pacific are also very relevant for the seasonal cycle and interannual variations (Magaña et al.,

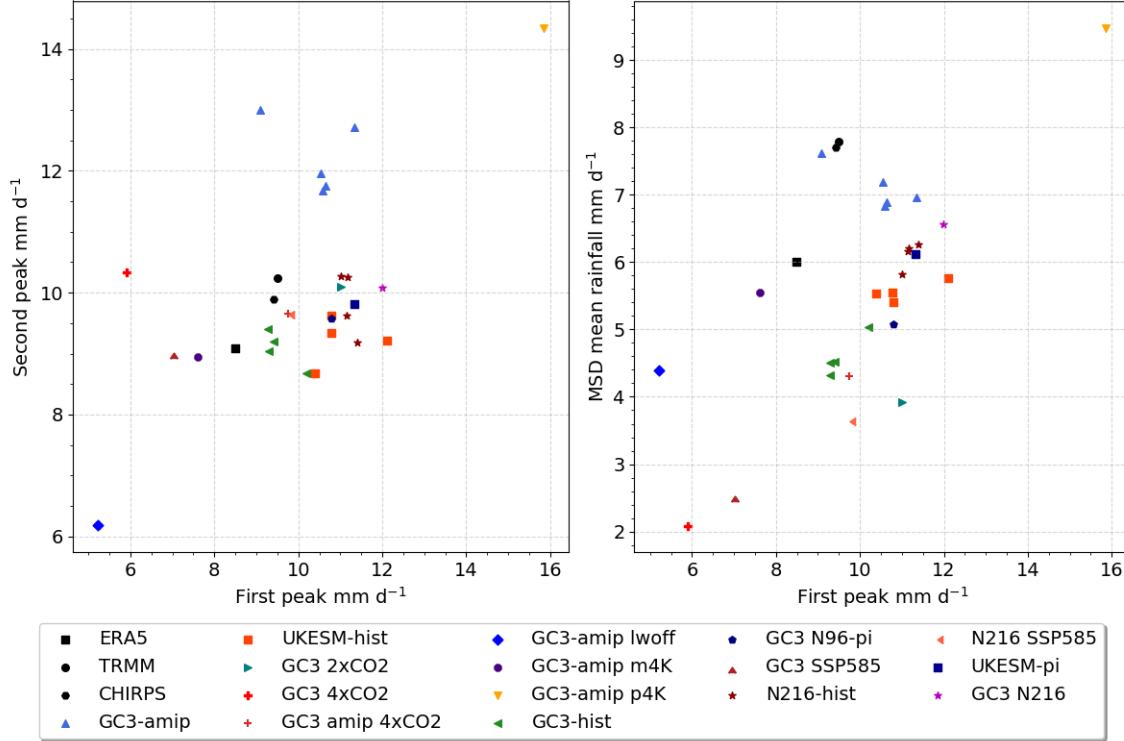


Figure 5.2: Scatter plots of the mean values of precipitation [mm day^{-1}] for TRMM, ERA5, CHIRPS and the experiments. (a) shows the scatter of the mean precipitation during the first peak versus precipitation during the second peak and (b) first peak versus the second peak.

1999; Amador, 2008; Straffon et al., 2019). Figures 5.5a, b show the seasonal cycle of SSTs in the EP and the Caribbean Sea. While the EP shows a maximum in SSTs in late May, during the early stages of the monsoon in Central America, the Caribbean SSTs peak in early fall, about five months later.

The Caribbean Low-level Jet (CLLJ) is a strong low-level easterly jet in the Caribbean Sea that peaks at the end of June (Figure 5.5e) at the 925 hPa level (Amador, 2008; Herrera et al., 2015; Maldonado et al., 2016). The CLLJ determines the moisture transport from the Caribbean Sea into the eastern Pacific across the Central American landmass as well as the northward moisture transport into the Gulf of Mexico and Florida (Muñoz et al., 2008; Hidalgo et al., 2015; Maldonado et al., 2016).

Figure 5.4 shows the distribution of rainfall in the different stages of boreal summer in different CMIP6 experiments and ERA-5. The main feature, the East Pacific ITCZ shows the maximum rainfall rates ($>15 \text{ mm day}^{-1}$ in the models) and strong mid-level ascent (-0.1 Pa s^{-1}). Prior to the MSD, rainfall extends from the easternmost Pacific ITCZ into the North American continent. Therefore, the positive bias during the first

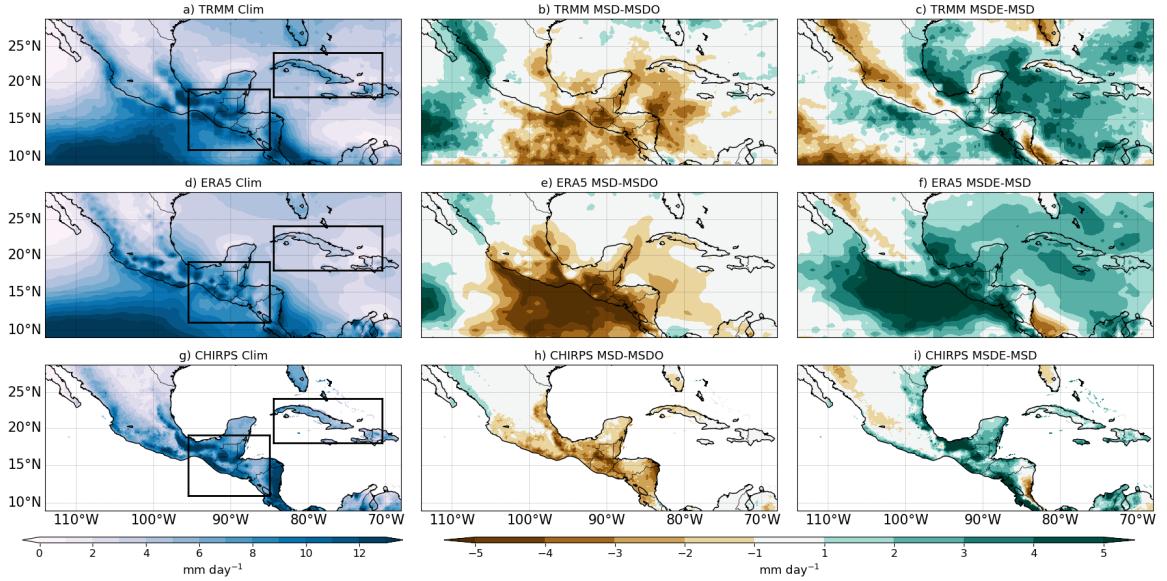


Figure 5.3: (a, d, g) Climatological JJAS rainfall and the difference between (b, e, h) the midsummer drought and the first peak periods and (c, f, i) between the second peak and the midsummer drought periods for (a-c) TRMM, (d-f) ERA5 and (g-i) CHIRPS.

peak over land is associated with the biased wetter EP ITCZ. However, during the MSD, rainfall decreases over land remaining only above 10 mm day^{-1} south west of the coastline in the models.

However, as shown in Figure 5.5a and as discussed for the radiative-convective feedback of Magaña et al. (1999), SSTs do not increase in the East Pacific in the late summer and the second increase in incoming shortwave is only modest in the reanalysis (Figure 5.5d).

However, SSTs in the easternmost Pacific do not increase after, during or at the end of the MSD (Figure 5.5a). In fact, the SSTs decrease with the second increase in deep convection and precipitation. The other hypothesis of this theory, referring to the incoming shortwave is also not consistent with observations, as the incoming shortwave only modestly increases during the midsummer (Figure 5.5d). There is perhaps a role for this modest increase in incoming shortwave, but the link to SSTs suggested by this theory does not agree with the reanalysis.

5.4 On the role of the East Pacific SSTs

Biases in the strength and position of the EP ITCZ in Global Coupled Models (GCMs) (Bellucci et al., 2010; Li and Xie, 2014; Schneider et al., 2014) are a major reason

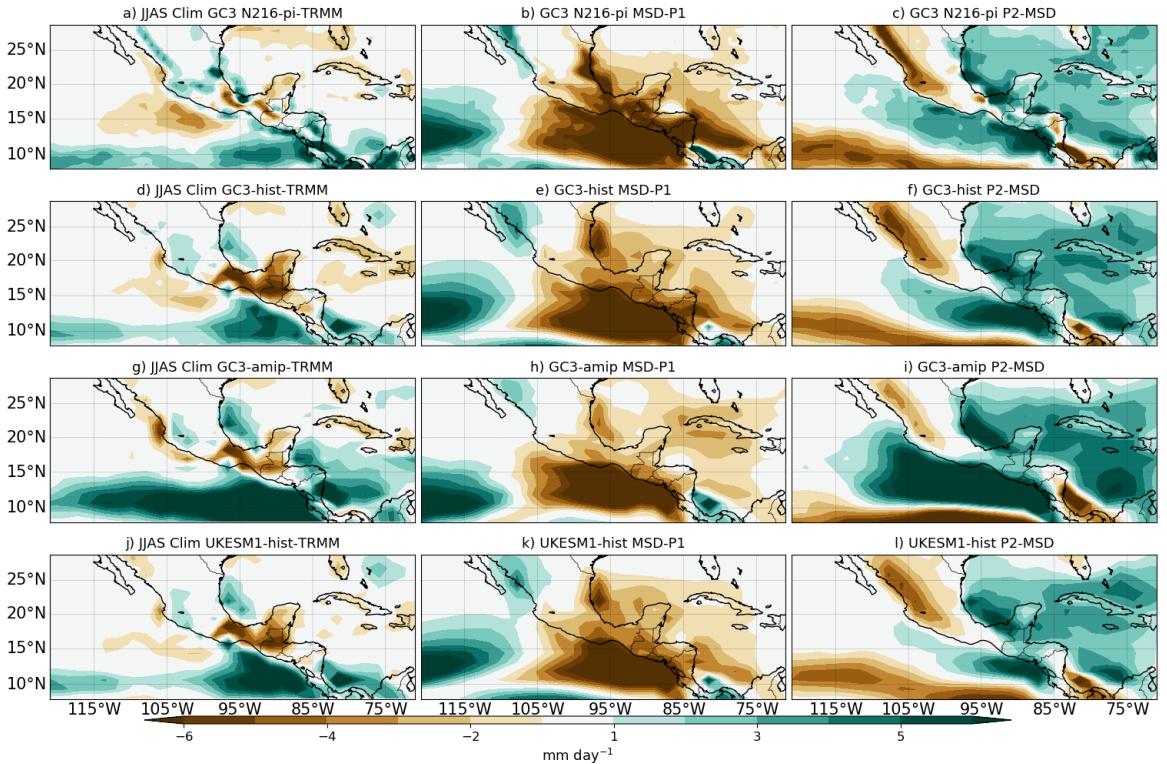


Figure 5.4: (a, d, g, j) JJAS model bias compared to TRMM and the difference between (b, e, h, k) the midsummer drought and the first peak periods and (c, f, i, l) between the second peak and the midsummer drought periods for four different simulations.

for biases in the model representation of rainfall in Central America (Rauscher et al., 2008).

Ryu and Hayhoe (2014) analyzed the performance of CMIP3 and CMIP5 models and found that the majority of CMIP5 models were unable to represent the total annual rainfall and the seasonal cycle of the MSD. Ryu and Hayhoe (2014) also finds that models that simulate a bimodal distribution of rainfall, HadGEM2-A for example, also show an accurate seasonal cycle of the NASH and the CLLJ. However, an exhaustive analysis as to whether these features are actually driving mechanisms for the MSD in GCMs as in observations is missing from the literature.

The CMIP6 Met Office models, HadGEM3 and UKESM1, are amongst the first models to simulate a bimodal regime in both Central America and Cuba (Figure 4.9a and 5.2). In Central America and southern Mexico, the models simulate a wetter-than-observed first peak of precipitation and a drier MSD period. The so-called second peak of precipitation found in late August is simulated in close agreement with TRMM, except in the AMIP experiment which has a far too strong second peak

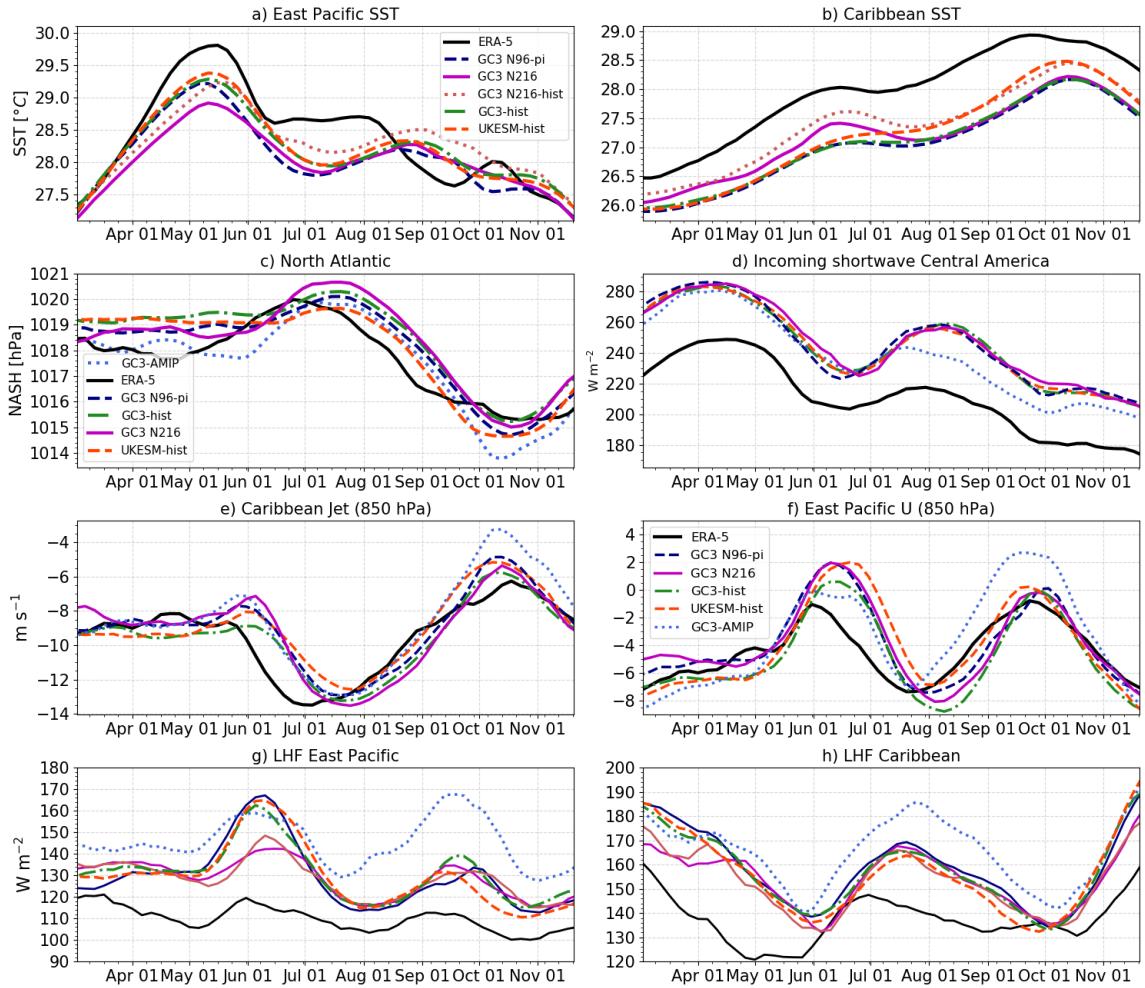


Figure 5.5: Pentad-mean seasonal cycle of indices associated with the MSD in Central America and the Caribbean.

mean precipitation rate.

5.5 On the role of the cloud-radiative feedbacks

The wetter EP ITCZ is a common feature of GCMs, including the Met Office models, which results from multiple biases in the radiative and convective schemes (Oueslati and Bellon, 2013; Li and Xie, 2014). In UKESM1 and HadGEM3 several biases exist in the radiative balance in the easternmost Pacific Ocean. A positive bias in incoming shortwave in Central America of about 15% and a cold SST bias in both East Pacific and Caribbean Sea SSTs are observed in Figure 5.5. Increased incoming shortwave but cooler SSTs require increased surface fluxes to maintain energy balance. These higher latent heat fluxes (LHFs) in the models in both basins (Figs. 5.5g, h) are

almost 40% larger than in ERA5 during the first peak of rainfall. The models also exhibit a larger seasonal cycle of the fluxes than the reanalysis. GC3 AMIP is the only simulation to also show a significantly positive bias in LHF during the second peak of rainfall in the EP but also at the end of MSD in the Caribbean Sea.

In all the model experiments, the ITCZ prior to the MSD period is stronger than in ERA5 by more than 5 mm day^{-1} , whereas after the MSD rainfall in the coupled models on the western coast of Central America agrees well with the ERA5. This analysis suggests that the biases shown in Figure 4.8 are mostly coming from the period prior to the MSD. The models reasonably simulate the decrease in rainfall during the MSD (Figure 5.4) followed by the second increase or peak. Note that GC3 AMIP, forced by very similar SSTs as ERA-5, simulated a much larger mean precipitation in the ITCZ during MSDE in contrast to the coupled models. This large positive bias in simulated rainfall in the East Pacific in GC3 AMIP corresponds to the larger than observed second peak observed in Figure 5.2a.

Composites prior to the onset of the MSD, during the MSD and after the MSDE were computed for several diagnostic variables. The periods were separated using the WT method to determine the dates of the MSDO and MSDE in ERA5 and the climate model output. Figure 5.6 shows the composite differences between the period of the MSD and of the two peaks in out-going longwave radiation (OLR) and vertical velocity ($\omega - 500$) at 500 hPa. The positive OLR and ω anomalies in the MSD-MSDO panels in southern Mexico and northern Central America are indicative of decreased height of convection and decreased ascent, in agreement with the MSD being the drier period. These positive anomalies in the continent are accompanied by negative OLR and $\omega - 500$ anomalies west of the continent, around 125°W .

The MSDE-MSD panels show the difference between the second peak of rainfall and the drier MSD period. Negative OLR and ω anomalies indicate stronger and higher convection over a wide region including the easternmost Pacific Ocean, southern Mexico, northern Central America Cuba and the Caribbean Sea. Note also the region of the North American Monsoon, on the northwest corner of Mexico and the southernmost US, as the MSD-MSDO difference suggests increased convective activity in the North American Monsoon region and MSDE-MSD the opposite.

Similarly, Figure 5.7 shows the low-level wind field during the three stages of the MSD. In ERA-5, prior to the MSD the wind flow in the Caribbean shows strong easterlies that flow into the Gulf of Mexico and southeastern US but very weak winds in the EP (see Figure 5.5f). During the MSD, the winds in the EP become modestly strong easterlies associated the easterly flow from the Caribbean Sea that crosses over

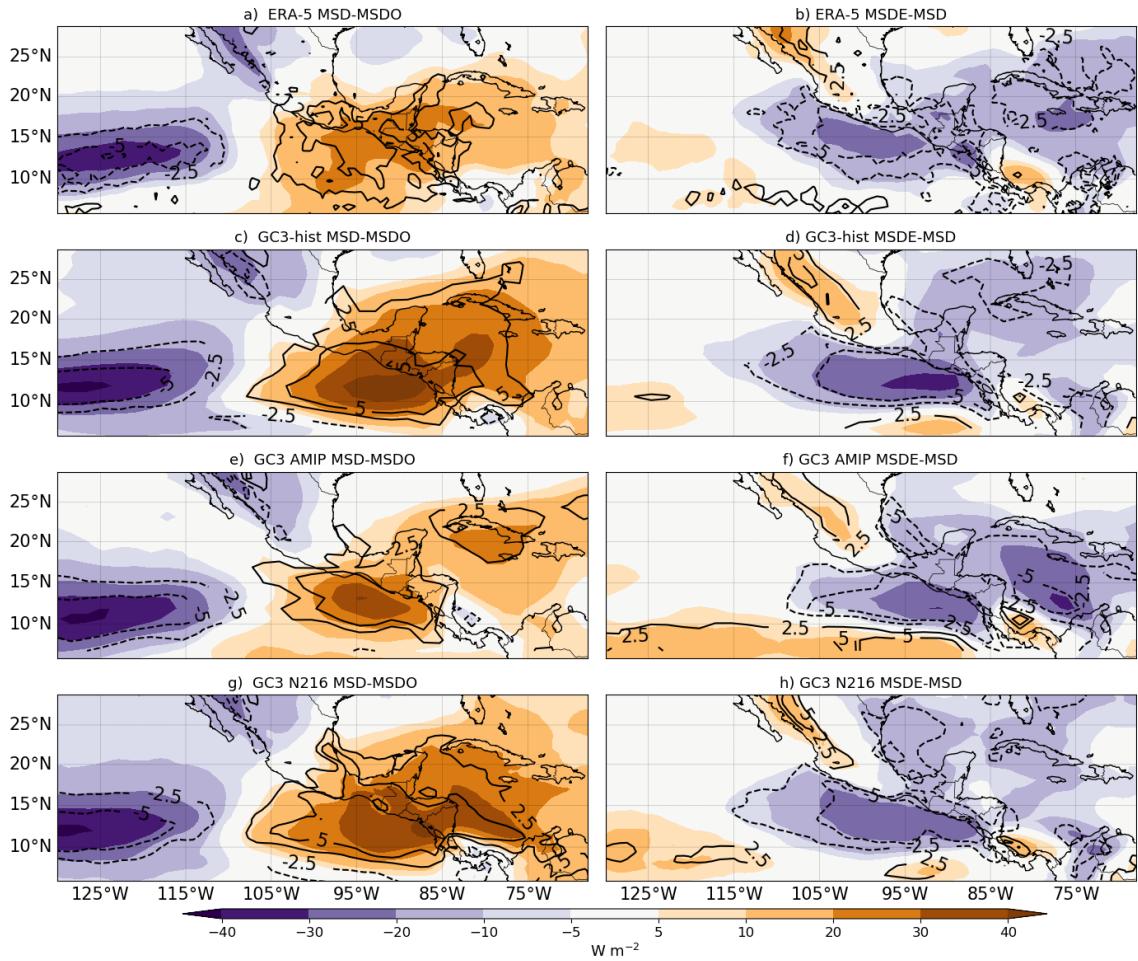


Figure 5.6: Out-going longwave radiation (OLR) [W m^{-2}] (shaded) and ω 500-hPa [$10^{-2} \text{ Pa s}^{-1}$] (line contours) differences between the MSD and MSDO and the MSDE and MSD.

Costa Rica and Nicaragua from the Caribbean Sea to the East Pacific. Note that the easterlies converge towards the region at 125°W where OLR and ω anomalies suggest increased ascent.

By the end of the MSD the easterlies in ERA5 weaken substantially on the western coast of Central America and in the Caribbean Sea. The simulations seem to generally reproduce the characteristics of the wind field with some differences worth mentioning. For instance, prior to the onset of the MSD, all the simulations show a modest westerly wind flow in the east Pacific at 10°N , which can also be seen in Figure 5.5f, which is not observed in ERA5. After the MSD ends, most simulations show a very weak westerly flow in the East Pacific, close to ERA5; however, GC3 AMIP shows a modest westerly wind converging towards the west coast of Nicaragua. This low-level convergence may

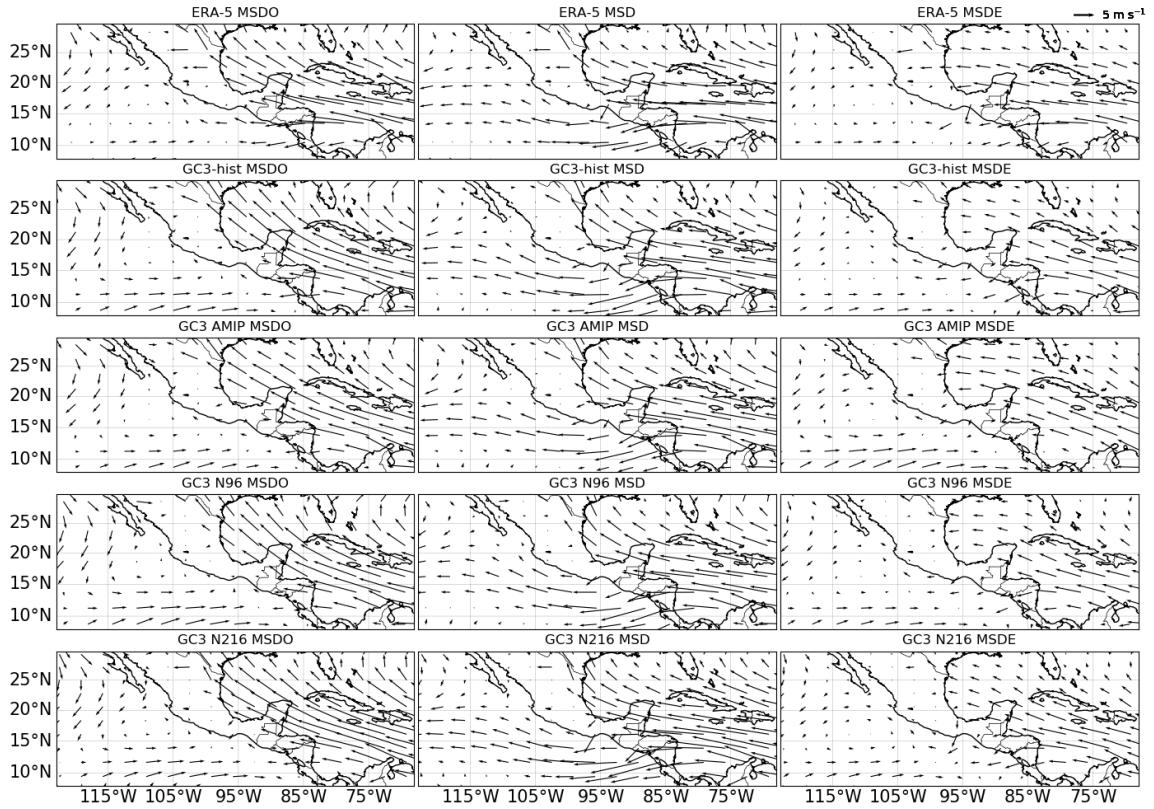


Figure 5.7: As in Figure 5.6 but showing wind vectors at the 850 hPa level.

be forcing the increased convective activity and precipitation during this time in GC3 AMIP.

The SSTs and incoming shortwave radiation are key elements for explain the seasonal cycle of the MSD, according to previous theories summarised in section ???. Figure 5.8 shows the corresponding SST and incoming shortwave anomalies during the different stages of the seasonal cycle. From the first peak to the MSD, a positive SST difference of +1.5 K in the Gulf of California and the western coast of the Baja California Peninsula is observed in reanalysis and the models. The differences appear as a sharp SST meridional gradient pattern around 115°W. During this stage, the incoming shortwave increases in Central America, which agrees with Figure 5.5d. Note the negative incoming shortwave differences west of Central America at 125°W, the region of negative OLR and ω -500 hPa anomalies where low-level winds converge, all of which supports the notion of increased convective activity that reduces incoming shortwave west of the continent. This feature was noted by Herrera et al. (2015).

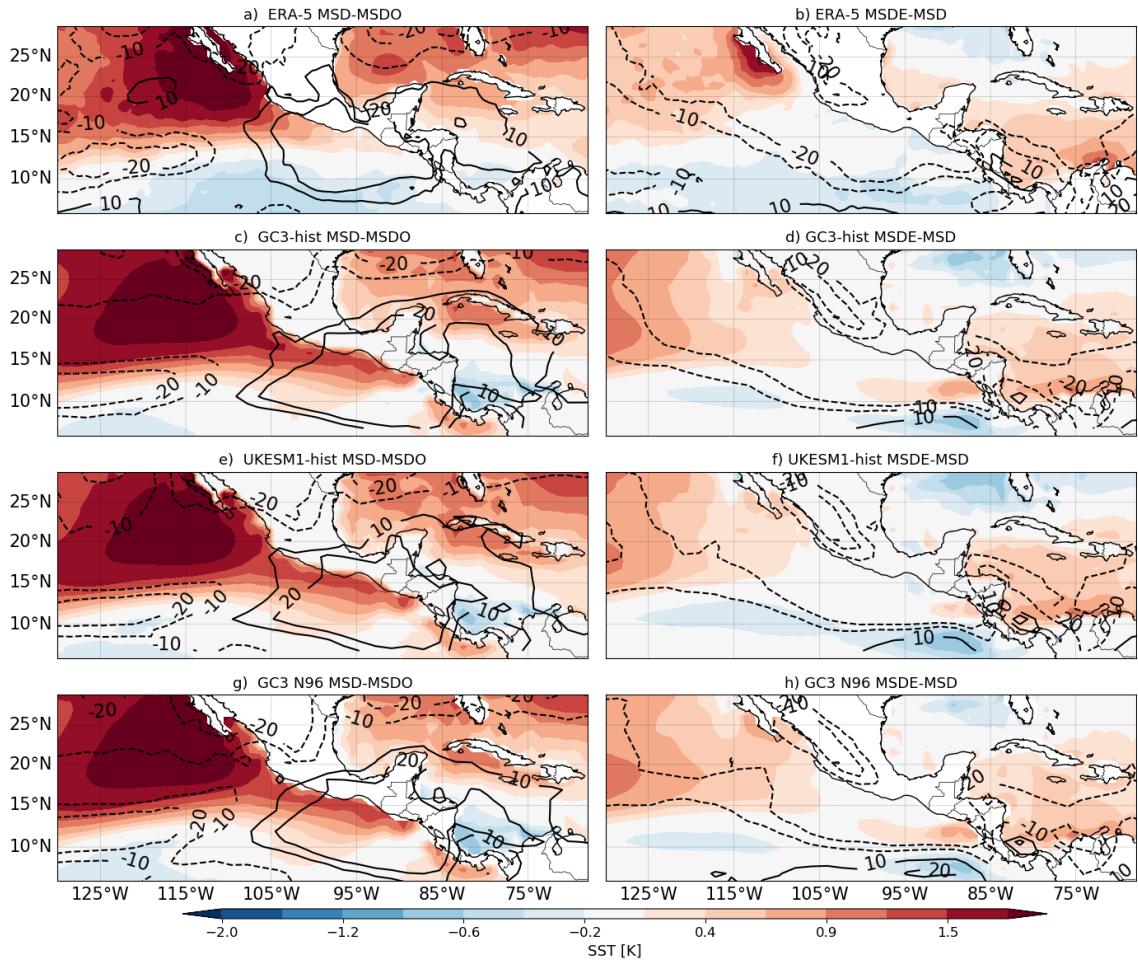


Figure 5.8: As in Figure 5.6 but the anomalies are shown for SSTs [K] (contours) and incoming shortwave radiation [W m^{-2}] at the surface (line-contours). Incoming shortwave is defined such as negative differences imply less incoming shortwave and positive anomalies represent more incoming shortwave at the surface.

5.5.1 On the role of the Caribbean Low-Level Jet

After the MSD, the western coast of the Baja California Peninsula continues to warm and the East Pacific continues to cool, in contrast to previous suggestions (Magaña et al., 1999; Magaña and Caetano, 2005; Herrera et al., 2015). Meanwhile, the Caribbean Sea warms by 1 K and the northern Gulf of Mexico slightly cools down. The incoming shortwave differences show a regional-scale decrease in incoming shortwave, as the summer draws to an end. These SST differences indicate that the meridional SST gradient in both the EP and Caribbean Sea and Gulf of Mexico is greatly modified during the stages of the MSD.

The main dynamical argument put forth to explain the MSD is centred around

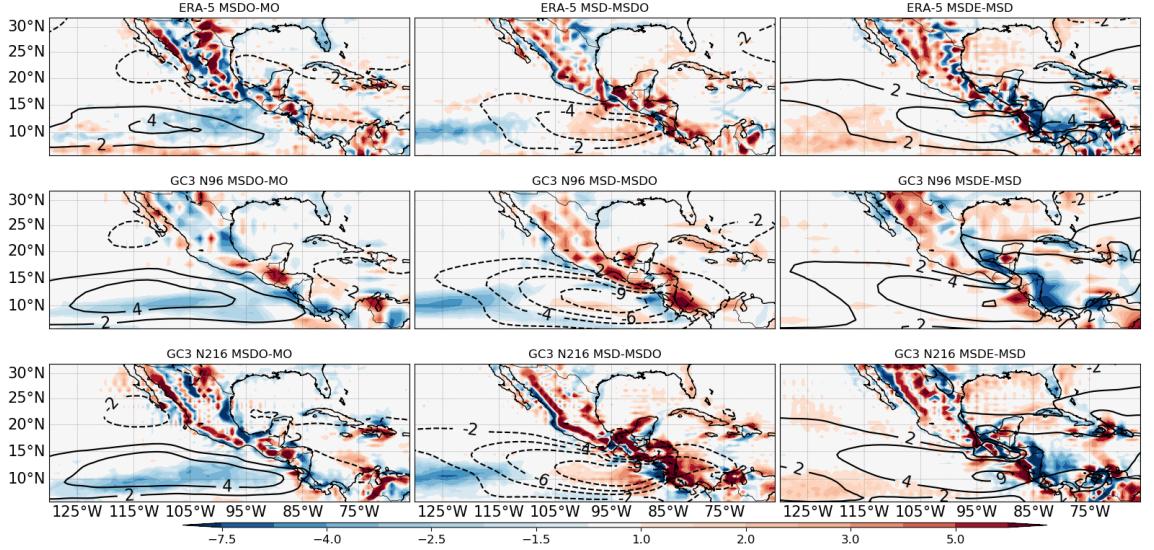


Figure 5.9: As in Figure 5.6 but showing in shading, moisture flux divergence $\nabla \cdot \vec{u}q$ at the 850 hPa level with units of $10^{-7} \text{ s}^{-1} \text{ kg / kg}$ and zonal wind anomalies (line contours) in m s^{-1} .

variations in the moisture flux convergence (MFC), argued to be driven by the Caribbean-Low Level Jet (see e.g. Gamble et al., 2008; Herrera et al., 2015; Martinez et al., 2019). The MFC and zonal wind variations in each stage of the MSD is shown in Figure 5.9 for ERA-5 and two simulations. The low-level MFC increases from monsoon onset (MO) to the first peak period (MSDO) in the EP. This anomaly in MFC corresponds to a region of positive zonal wind anomalies indicative of weaker easterly flow. This zonal wind anomaly from MSD to MSDO is much stronger in the models. The MSD-MSDO difference shows a strong positive MFC anomaly across southern Mexico and most of Central America.

In turn, the MFC anomalies associated with the end of the drier period, observed as the MSDE-MSD anomalies, show negative values, suggesting increased moisture flux, over southern Mexico and northern Central America. Increased moisture flux during the transition from the MSD to the second peak agrees well with the precipitation differences during these periods. The MSDE-MSD zonal wind anomalies in the EP show positive zonal wind anomalies, suggesting a weakened easterly wind flow (see also Fig. 5.7).

The MSD in Central America and southern Mexico has been strongly linked to the strengthening of the CLLJ (Herrera et al., 2015). The maximum zonal wind observed in the CLLJ is found at the very end of July (Fig. 5.5e), synchronized with the start of the MSD. The zonal wind anomalies in the MSD-MSDO panels

in Figure 5.9 show that easterlies in the Caribbean Sea do not strengthen by more than 2 m s^{-1} from the first peak to the MSD. Only in the models is there a modest negative anomaly at the westernmost Caribbean Sea. In other words, while the peak of the climatological CLLJ coincides with the climatological timing of the onset of the MSD, these composite analyses constructed by more specifically separating the MSD periods does not show relevant variations in the zonal wind of the Caribbean Sea. The drier MSD period does coincide with stronger easterly flow over the eastern Pacific, which may be associated with the weaker MFC over land.

5.5.2 A look at the MSD through the lense of the moist static energy budget

5.6 Summary and discussion

The midsummer drought is a prominent feature of the seasonal cycle of rainfall of southern Mexico, northern Central America and the Caribbean. The average 20% decrease during the midsummer compared to the wetter periods of early and late summer is a rare feature of monsoon regions that has important implications for agriculture and water management (Hellin et al., 2017; de Sousa et al., 2018; Harvey et al., 2018).

Climate predictions of the MSD, particularly those concerning whether this "drought" will become more pronounced in the following years, are not trustworthy because of several reasons. One factor is the current limitation in the understanding of the physical processes that cause the MSD (section ??) as debate still exists over which large or regional-scale processes are most important to explain the increases and decreases of precipitation over intraseasonal time-scales. Secondly, methods used to diagnose the timing and strength of the MSD typically deal with monthly-scale metrics, which would obscure subtle trends and processes that have an effect on shorter time-scales. Also relevant is the fact that climate models used to produce the predictions show significant biases in the EP ITCZ and the seasonal cycle of rainfall in the region, in fact, most CMIP3 and CMIP5 models did not show a bimodal signature in the seasonal cycle. Models that do not have a climatological MSD cannot provide a prediction for this regime in future climate.

For these reasons, this section analysed the CMIP6 simulations from the Met Office models, UKESM1 and HadGEM3, aiming to understand the causes of the biases in the seasonal cycle. Furthermore, these models are better compared to CMIP3 and CMIP5 cohorts since UKESM1 and HadGEM3 actually simulate a bimodal precipitation

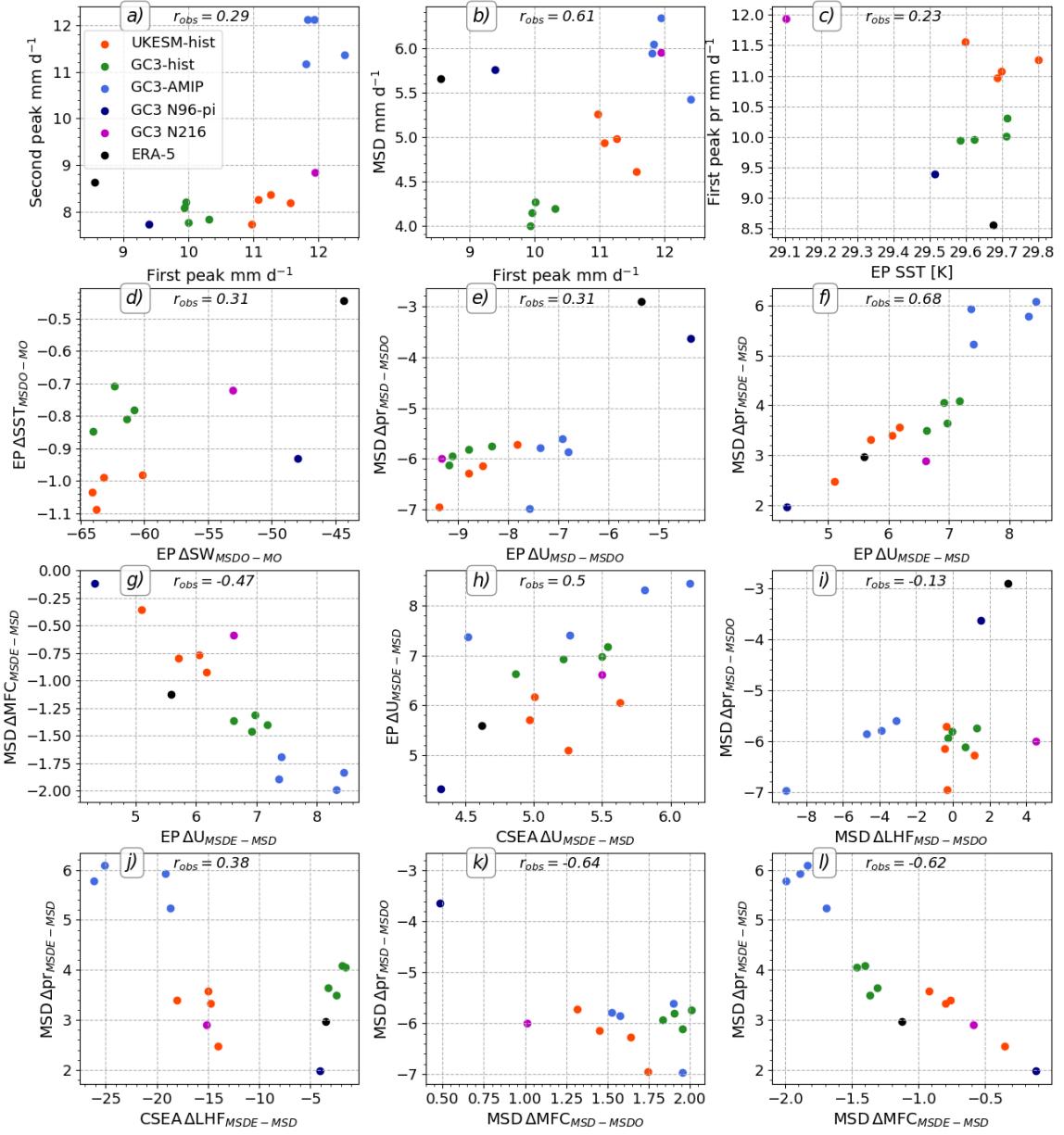


Figure 5.10: Scatter plot of the (a, b) area-averaged precipitation over land (Box in Figure 4.8) during the different stages of the MSD. (c) scatter of the East Pacific SSTs against the precipitation over land during the first peak period. (d-l) show the scatter differences in several variables between the different stages of onset of the MSD (MSDO), the drier MSD and the end of the MSDE. The differences are shown for area-averaged quantities in the East Pacific (EP), the Caribbean Sea (CSEA) and overland (MSD) as above. The units for ΔU are $[\text{m s}^{-1}]$, ΔMFC $[10^{-11} \text{ s}^{-1}]$, ΔSW and ΔLFH $[\text{W m}^{-2}]$ and $\Delta \text{pr mm d}^{-1}$. The Pearson correlation coefficient for the 38 yr of reanalysis or observations (r_{obs}) is shown for each panel.

regime in these regions. The purpose of this investigation is to use these climate models to better diagnose the relevant biases for the representation of the MSD but also understand the processes that these models are capturing leading to the MSD, in order to, hopefully, also highlight the dynamics of the MSD in general.

The wavelet transform method was developed to determine the pentads of onset and end of the MSD. For instance, Figures 5.10a,b show the scatter of the mean precipitation during the first peak against second peak and first peak against MSD in all the simulations and ERA5. The magnitude of the first and second peaks appear to be unrelated in these models and in observations, which would suggest that the processes driving each peak are not exactly the same. Similarly, composite analysis of various diagnostics during the different stages of the seasonal cycle was done, for instance, OLR composites showed that the MSD is not a local feature in a small region of southern Mexico but extends throughout a wide range of North America, from central Mexico through Belize, Guatemala, El Salvador, Honduras, Nicaragua, and northern Costa Rica.

This composite approach also allowed to test previously proposed hypotheses by analysing the differences between model experiments and the observed variability in the characteristics of the precipitation at each stage of the MSD. For example, Magaña et al. (1999) proposed a mechanism that explains the MSD through SST-cloud feedbacks. In this hypothesis, shortwave, SSTs and precipitation are strongly coupled in the EP Ocean. The first peak of precipitation in southern Mexico and Central America would then be associated with the EP SSTs prior to the onset of rainfall. Figure 5.10c shows that EP SSTs prior to onset do not explain the inter-model differences in the magnitude of the first peak nor do they show a strong relationship in the observed interannual variability of the first peak mean precipitation. Similarly, Figure 5.10d shows that surface incoming shortwave variations are only weakly related to SSTs variations in the EP, in both models and reanalysis, during the first peak period.

The feedback mechanism also suggests that the second peak is a result of a second increase in surface incoming shortwave that occurs as cloud cover decreases during the drier MSD. This increase in incoming shortwave then increases EP SSTs and thus increasing convective activity. Although the incoming shortwave does show a bimodal behaviour (Figure 5.5d), the SSTs in the East Pacific do not increase during the MSD period, but in fact cool during the end of the MSD. Furthermore, as in Figure 5.10d, variations in incoming shortwave were not strongly related to SST changes in any of the stages of the MSD (not shown). This suggests that the SSTs are not only dependent on the incoming shortwave in both models and reanalysis.

The low-level winds (Figure 5.7) show notable changes between the onset of the MSD (MSDO), the MSD and the end of the MSD (MSDE). Weak westerlies in the EP are found during the wetter periods but the zonal wind becomes a modest easterly flow during the drier MSD period. The MSDO appears to be synchronized with the strengthening of the Caribbean Low-Level Jet (Fig. 5.5e). During the MSD, the strong zonal flow in the Caribbean crosses Central America into the central-eastern Pacific. This easterly flow during the MSD converges to 125°W in the EP Ocean, a region that also shows increased ascent during the MSD.

Figure 5.10e, f show the relationships between the zonal flow in the EP Ocean and precipitation in southern Mexico and Central America. The changes in the wind flow between the first and the MSD are not related to the drying response over land during the same period. However, the differences between the second peak and the MSD in the wind flow and precipitation show a strong relationship both in observed interannual variability as well as in the model spread. Simulations with a stronger EP zonal wind anomaly show the strongest increment in precipitation over land. The zonal wind change in the EP from the MSD to the second peak period is also modestly related to the MFC over the continent (Fig. 5.10g) with weaker easterly winds in the EP associated with more convergence over land in the models and reanalysis.

The easterly flow in the EP has been associated with the strength of the CLLJ (Herrera et al., 2015). The zonal wind changes in the MSDE-MSD difference in the EP shows a modest linear relationship with the zonal flow in the Caribbean Sea (Fig. 5.10h). During the other periods, the relationship between the CLLJ and the EP zonal component of the wind is even weaker in both models and observations (not shown).

A potentially relevant bias found in the models was stronger-than-observed surface latent heat fluxes (LHF) (Figure 5.5g, h) compared to the reanalysis. Changes in the surface energy balance and the surface temperature in historical versus pre industrial control simulations may also be responsible for the precipitation differences between these experiments. However, the variations in the LHFs, both MSD-MSDO and MSDE-MSD either in the Caribbean Sea or over land (Figure 5.10i,j) are not related to precipitation over land.

The main factor associated with the precipitation variations in the seasonal cycle appears to be the low-level moisture flux convergence (MFC) (Figure 5.10k, l). The variations in the MFC over land explain intermodel differences and observed interannual variability in precipitation, particularly in the positive rainfall increment from

the MSD to the second peak. From the first peak to the MSD, moisture flux decreases and increases again from the MSD to the second peak.

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