The American Monsoon System: variability and teleconnections

Jorge Luis García Franco

Wadham College
University of Oxford

A thesis submitted for the degree of
Doctor of Philosophy

Michaelmas 2020

Abstract

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Contents

List of Figures

List of Abbreviations

1-D, 2-D . . . One- or two-dimensional, referring in this thesis to spatial dimensions in an image.

Otter . . . . . One of the finest of water mammals.

Hedgehog . . . Quite a nice prickly friend.

Neque porro quisquam est qui dolorem ipsum quia dolor sit amet, consectetur, adipisci velit...

There is no one who loves pain itself, who seeks after it and wants to have it, simply because it is pain...

— Cicero's *de Finibus Bonorum et Malorum*

1 Introduction

Contents

1.1 Motivation

The American Monsoon System (AMS) provides the majority of rainfall for the large regions in Latin America and southwestern United States. Climate variability and teleconnections to this monsoon system can impact the population through changes in extreme precipitation, the timings of the monsoon or the overall rainfall during the rainy or the dry seasons causing floods or droughts.

General circulation models (GCMs) have been used to provide climate projections of future climate in the AMS. However, GCMs may also be used to understand physical mechanisms associated with climate variability and teleconnections.

This thesis focuses on the American Monsoon System and the outstanding questions regarding the climate variablity and teleconnections affecting this monsoon.

1.2 Contribution

Chapter 3 evaluates two state-of-the-art CMIP6 models for their representation of the monsoon system. In general, the models show a good representation of the seasonal cycle as they are able to simulated detail aspects such as the Midsummer drought. ENSO teleconnections in these models appear to be non-linear, as are the observations. Chapter 4 provides a method that is able to better characterise the MSD timings and strenghths, 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.

Alles Gescheite ist schon gedacht worden.
Man muss nur versuchen, es noch einmal zu denken.

All intelligent thoughts have already been thought;
what is necessary is only to try to think them again.

— Johann Wolfgang von Goethe **?**

2
Background

Contents

This chapter summarises the main aspects of the tropical circulation and of the global monsoon. Then, the relevant background of the characteristics of American monsoon system and outstanding issues is given.

2.1 The tropical circulation and the global monsoon

Tropical climate is characterized by the strong incoming solar insolation year-round which makes the tropics the warmest region of the planet. The differential heating between the tropics and higher latitudes drives a meridional transport of energy by the atmosphere. The tropical oceans that receive such a strong solar insolation exhibit strong evaporative fluxes. The in combination with the differential heating of land over ocean modulate the tropical circulation. This tropical circulation can be described to a first order through the zonal and meridional circulations known as the Hadley and Walker cells.

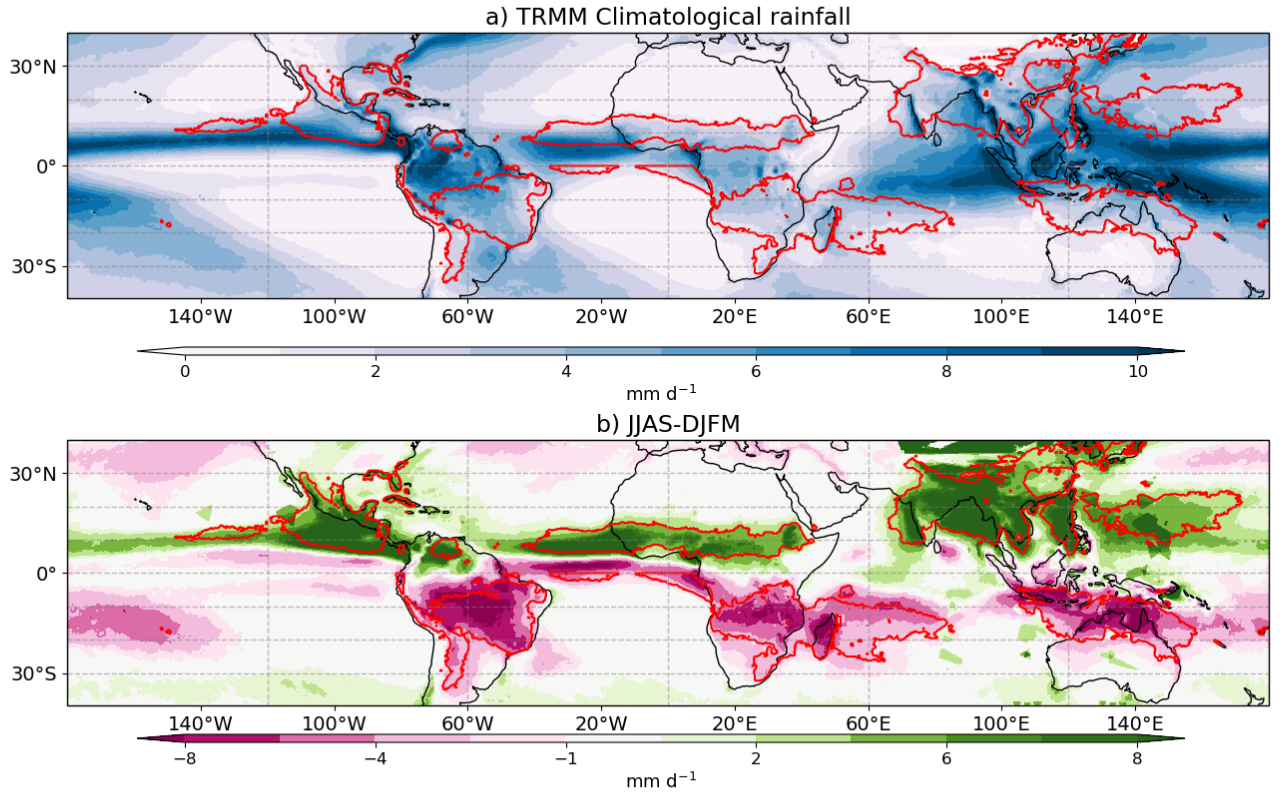

Figure 2.1: a) Climatological mean annual rainfall rates in the tropics in the TRMM dataset (1999-2018). b) The mean rainfall rate difference between boreal summer (JJAS) and austral summer (DJFM). The red contours highlight the regions where the summer rainfall amount accounts for more than 55% of the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

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). 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(Bjerknes, 1969; Gill, 1980). The dynamic and thermodynamic effects of the location and strength of convection in the Walker circulation have strong impacts across all the tropics and also the extratropics (Cai et al., 2019).

The Inter-tropical Convergence Zone (ITCZ) is a band of precipitation that migrates meridionally with the seasons (Schneider et al., 2014). The ITCZ is arguably one of the most relevant features of tropical climate because tof the strong influence on the low- and upper-level circulation associated with ITCZ and the fact that the vast majority of rainfall in the tropics is due to convection in the ITCZ. The ITCZ is characterized by a strong 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 traditional view of a monsoon is that of a large-scale land-sea breeze associated with the different heatings of the land and the ocean that force a seasonal reversal of the winds. However, this framework has recently been challenged by the notion that monsoons are simply a manifestation of the ITCZ migration (Gadgil, 2018). This has created the concept of the global monsoon as a phenomenon that exists all across the continental tropics (Zhou et al., 2016; Gadgil, 2018).

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.

2.2 The American Monsoon System

The AMS is typically subdivided into the North and South American monsoon systems (Vera et al., 2006). The North American Monsoon is the main source of rainfall in south-western North America, 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). A key aspect of this monsoon is the moisture advected by the low-level flow from the

Gulf of California and the East Pacific Ocean and to a lesser extent the moisture mixed in the mid-troposphere from the Caribbean Sea and Gulf of Mexico (e.g Stensrud et al., 1997; Pascale and Bordoni, 2016; 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).

2.3 Review of the Midsummer drought

The seasonality of precipitation in northwestern Central America and southern Mexico fits the new definition of a monsoon characterized by the sharp contrast between the wet and dry seasons. 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).

These variations of precipitation are 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). This feature of the seasonal cycle is most commonly referred to in the literature as Midsummer drought (MSD). Farmers in Central America who are subject to climatic stress due to droughts, have already perceived and adapted to changes in the characteristics of the rainy season, such as the timing and strength of the midsummer drought (Hellin et al., 2017; de Sousa et al., 2018; Harvey et al., 2018), but it is unclear whether these perceived changes are a real trend in the observations (Anderson et al., 2019).

Observations have characterised this feature in the climatological precipitation of several regions of Mexico, El Salvador, Belize, Guatemala, 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 and the Caribbean. 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; Ryu and Hayhoe, 2014; 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. Dynamical or the thermodynamical mechanisms have been put forth and different roles have been proposed for the Atlantic and the East Pacific Oceans (e.g. Magaña and Caetano, 2005; Gamble et al., 2008; Herrera et al., 2015).

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.

Magaña et al. (1999) and Magaña and Caetano (2005) proposed a mechanism driven by radiative-convective feedbacks between the East Pacific 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 peak in SSTs during May (Figure 4.4a) triggers evaporation and deep convection in the East Pacific ITCZ and Central America (Figure 4.1). The high convective clouds produce a radiative cooling effect at the surface due to decreased incoming shortwave radiation (Figure 4.4d). 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 feedback effect caused by decreased frequency of tall convective clouds during July and August, as convective activity decreased during the MSD, which reduces the cooling effect of the clouds and increases incoming shortwave, SSTs and surface fluxes, eventually increasing precipitation (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 determines 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 moisture flow from the Caribbean Sea to the continent and the easternmost Pacific Ocean is another active topic of research (Hidalgo et al., 2015; Corrales-Suastegui et al., 2020). The effect of the SST gradient between the easternmost Pacific Ocean and the Caribbean Sea on the moisture flow across the continent is a relevant for variability on interannual time-scales (Martinez et al., 2020). 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 shifting 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) 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 double crossing of the Intertropical Convergence Zone (ITCZ), 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 Niño Southern Oscillation: impacts to the American monsoon system

ENSO is the leading mode of interannual variability in the tropics which produces changes in the atmospheric circulation that impact most regions on the planet. ENSO has motivated extensive research using GCMs to understand the mechanisms related the origin of ENSO (Christensen et al., 2017), the feedbacks and processes that phase-lock the phenomena (Neelin et al., 1998), as well as how will ENSO characteristics change in the future (Cai et al., 2015a; Santoso et al., 2017).

ENSO is the oscillation from a mean state of the Walker circulation and the SST gradient. El Niño was initially the term to refer to the SST changes and the Southern Oscillation

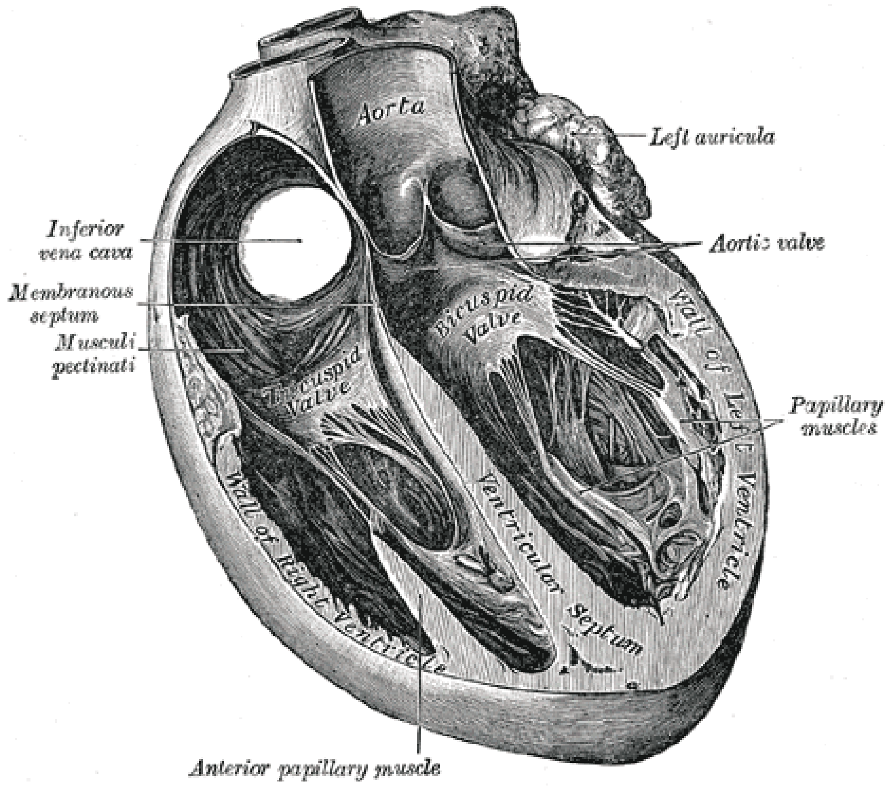

Figure 2.2: Four-chamber illustration of the human heart. Clockwise from upper-left: right atrium, left atrium, left ventricle, right ventricle.

(SO) is associated with changes in the zonal gradient of MSLP. Combined, El Niño and the SO compose the ENSO phenomenon.

The most prominent characteristic of ENSO is the effect that zonal gradients of MSLP and SST have on the location and strength of deep convection in the equatorial Pacific (Trenberth, 1997; Neelin et al., 1998). As ENSO shifts from the positive to the negative phase, deep convection shifts along the equatorial Pacific, generating precipitation anomalies on inter-annual timescales (Neelin et al., 1998; Wang and Picaut, 2004). In other words, ENSO poses a strong control on the location and strength of the Walker circulation which then produces effects in remote regions of the world where the Walker circulation is important, *e.g.*, in West Africa (Ropelewski and Halpert, 1986, 1987) or South America (Sulca et al., 2018).

The period of ENSO of 2-7 years (Neelin et al., 1998; Wang and Picaut, 2004) was poorly represented in CMIP3 and CMIP5 models (Guilyardi et al., 2009), particularly by models that had much stronger power spectrums than the observed.

2.5 Stratosphere-Troposphere Coupling in the Tropics

The QBO is the dominant mode of tropical stratospheric variability which modulates the characteristics of the tropopause in the tropics (Baldwin et al., 2001; Tegtmeier et al., 2020) such as height and cold point temperature. Observations have shown an influence of the QBO on deep convective features such as monsoons (Giorgetta et al., 1999), tropical cyclones (Chan, 1995) and the Walker circulation (Collimore et al., 2003). More recently a link between the QBO and the Madden-Julian Oscillation (MJO) was discovered (Son et al., 2017) and motivated extensive research (see e.g. Lee and Klingaman, 2018; Wang et al., 2019; Martin et al., 2020).

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 was 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 cumulunimbus clouds.

The use of climate models to understand these tropical teleconnections of the QBO has proven difficult due to biases in both the MJO and the QBO representations. State-of-the-art CMIP6 models struggle to reproduce several of the characteristics of the QBO (Richter et al., 2020). For instance, weaker temperature QBO signals in the lowermost tropical stratosphere in the models, e.g., compare the QBOE-W difference plot in Figure **??**. The weaker temperature signal may be key for possible biases in the tropical QBO links discussed above, such as the the QBO-MJO link which is missing from CMIP6 models (Kim et al., 2020) and from seasonal prediction forecast models (Wang et al., 2019; Martin et al., 2020).

3
Data and methods

Contents

3.1 Observations and reanalysis data

Table 3.1 summarises relevant information of the observations and reanalysis datasets used in this study. In short, surface and satellite observations were used where available, whereas other metrics were taken from reanalysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF): ERA-5, downloaded from `https://climate.copernicus.eu/climate-reanalysis`. Four different precipitation datasets are used.

The Tropical Rainfall Measurement Mission (TRMM) dataset is a multi-satellite multi-sensor product that in some versions is calibrated with gauge analyses (Huffman et al., 2007). A set of microwave 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. 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 measurements onboard geosynchronous earth orbit satellites. Other sources of information include a radar

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).

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

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. 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.

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, we use TRMM as our best estimate for the spatial and temporal characteristics of the AMS rainfall. However, the period covered by TRMM (1998-2018) is too short to analyse statistically robust teleconnections or variability, so we use GPCP, GPCC and CHIRPS 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).

3.2 Model data

The 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 most substantial change from the version used in CMIP5 (HadGEM2-AO) is the inclusion of the new GC configuration 3.1 (Walters et al., 2019) with the updated 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; for details see Williams et al. (2018) and Kuhlbrodt et al. (2018). The GC3 model was run for CMIP6 deck experiments

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

with two horizontal resolutions: a low resolution configuration, labelled as N96, with an atmospheric resolution of 1.875°x1.25° and a 1° resolution in the ocean model and a medium resolution configuration, labelled N216, with atmospheric resolutions of 0.83°x 0.56° 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.

Alles Gescheite ist schon gedacht worden.
Man muss nur versuchen, es noch einmal zu denken.

All intelligent thoughts have already been thought;
what is necessary is only to try to think them again.

— Johann Wolfgang von Goethe **?**

4

On the dynamical and thermodynamical mechanisms of the MSD in the Met Office CMIP6 models

Contents

4.1 Climatological features

Figure 4.1 shows the pentad-mean seasonal cycle of precipitation in Central America and the Caribbean. The seasonal cycle in both regions follows that of a monsoon, i.e., a dry winter and a wet summer season. In the first region (Figures 4.1a, b), two precipitation maxima, in June and September, are separated by a decrease in precipitation during July and August, *i.e* the MSD. In Central America, 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. The two peak structure in the Caribbean (Figures 4.1c, d) is characterised by two peaks in May and October with a four-month drier period in between the two peaks (e.g. Giannini et al., 2000; Gamble et al., 2008; Angeles et al., 2010). In Cuba, the difference between the first peak (June 01) to the driest pentad of the MSD (Aug 01) is of about 3 mm day^{-1} in the TRMM dataset.

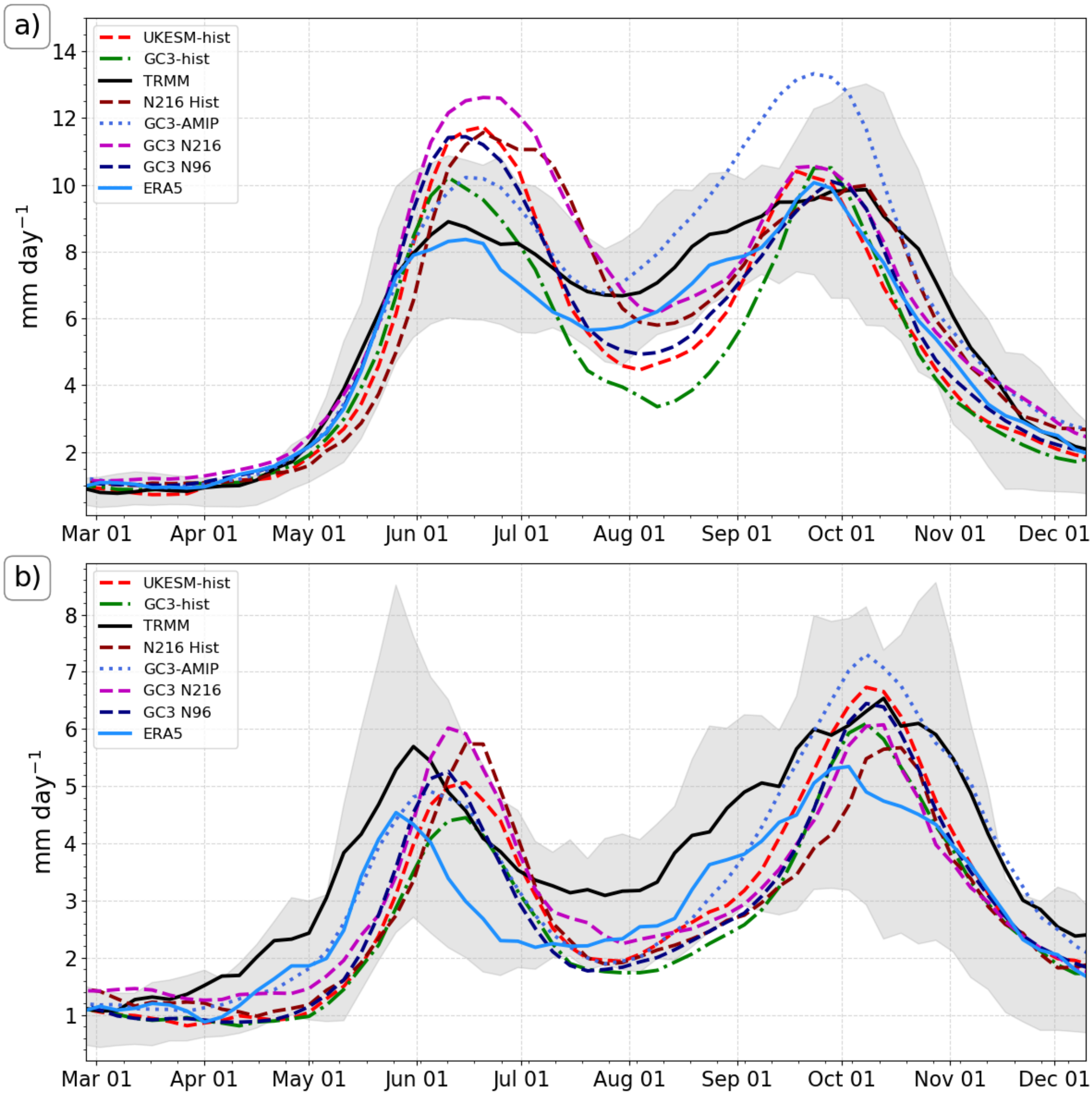

Figure 4.1: Pentad-mean precipitation in (a) southern Mexico and northern Central America and (b) Cuba. Shading shows uncertainty obtained by bootstrapping the interannual variability of the TRMM dataset.

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., 1999; Amador, 2008; Straffon et al., 2019). Figures 4.4a, 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

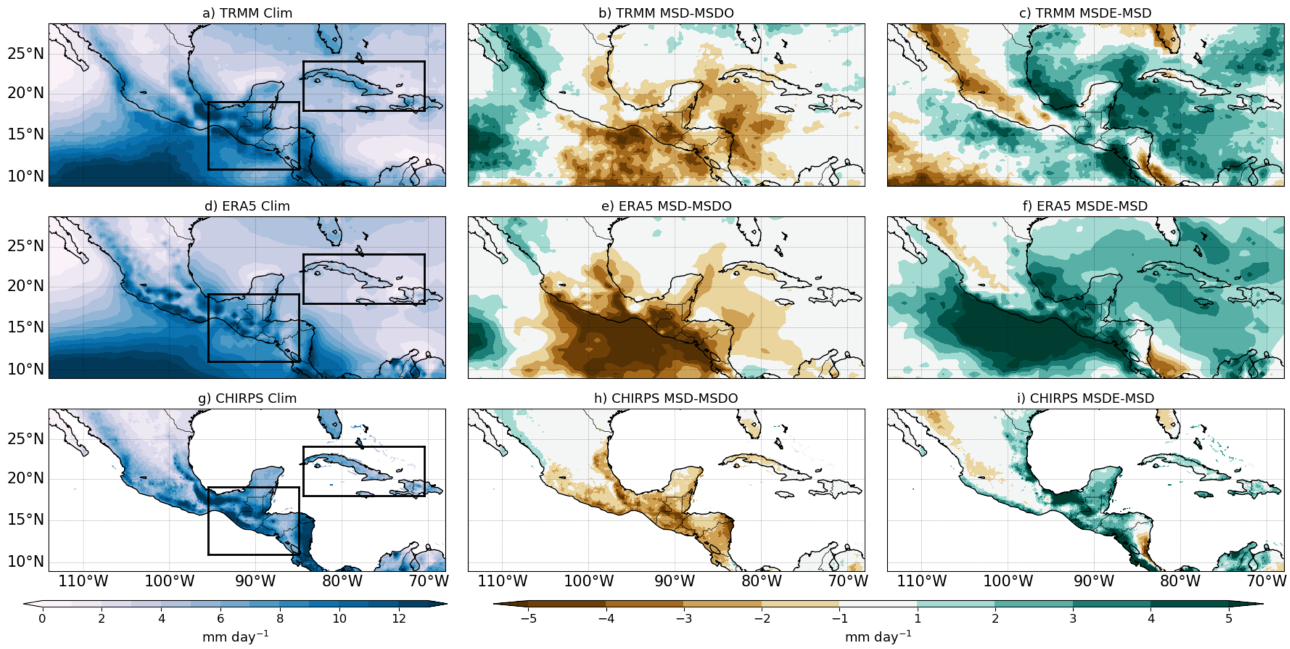

Figure 4.2: (a, d, g) Climatological JJAS rainfall and the difference between (b, e, h) the midsummer drought and the first peak periods and (c, f, i) between the second peak and the midsummer drought periods for (a-c) TRMM, (d-f) ERA5 and (g-i) CHIRPS.

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 4.4e) 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).

However, as shown in Figure 4.4a 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 4.4d).

However, SSTs in the easternmost Pacific do not increase after, during or at the end of the MSD (Figure 4.4a). 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 4.4d). 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.

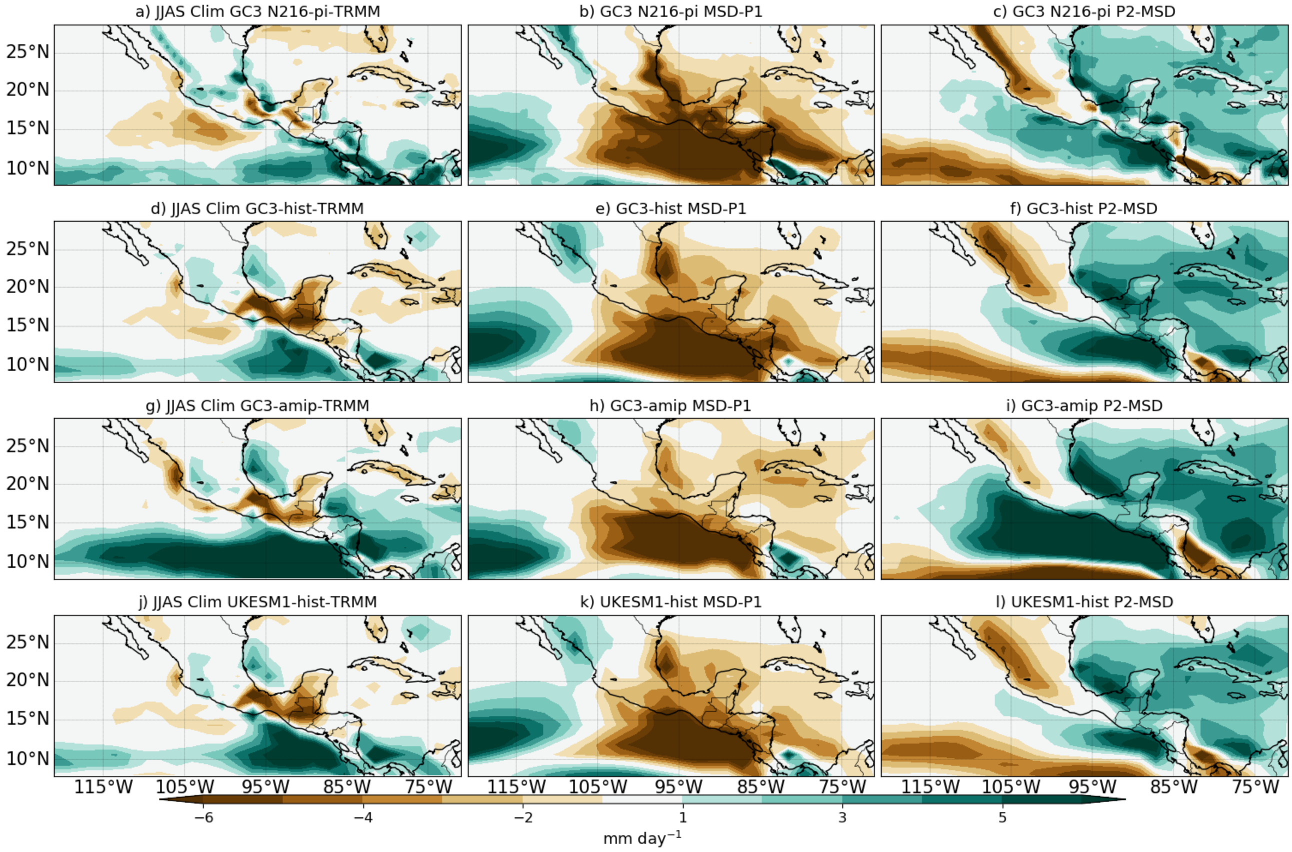

Figure 4.3: (a, d, g, j) JJAS model bias compared to TRMM and the difference between (b, e, h, k) the midsummer drought and the first peak periods and (c, f, i, l) between the second peak and the midsummer drought periods for four different simulations.

4.1.1 On the mechanisms of the MSD in the UK Met Office models

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 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 **??**a and 4.1). In Central America and southern Mexico, the models simulate a wetter-than-observed first peak

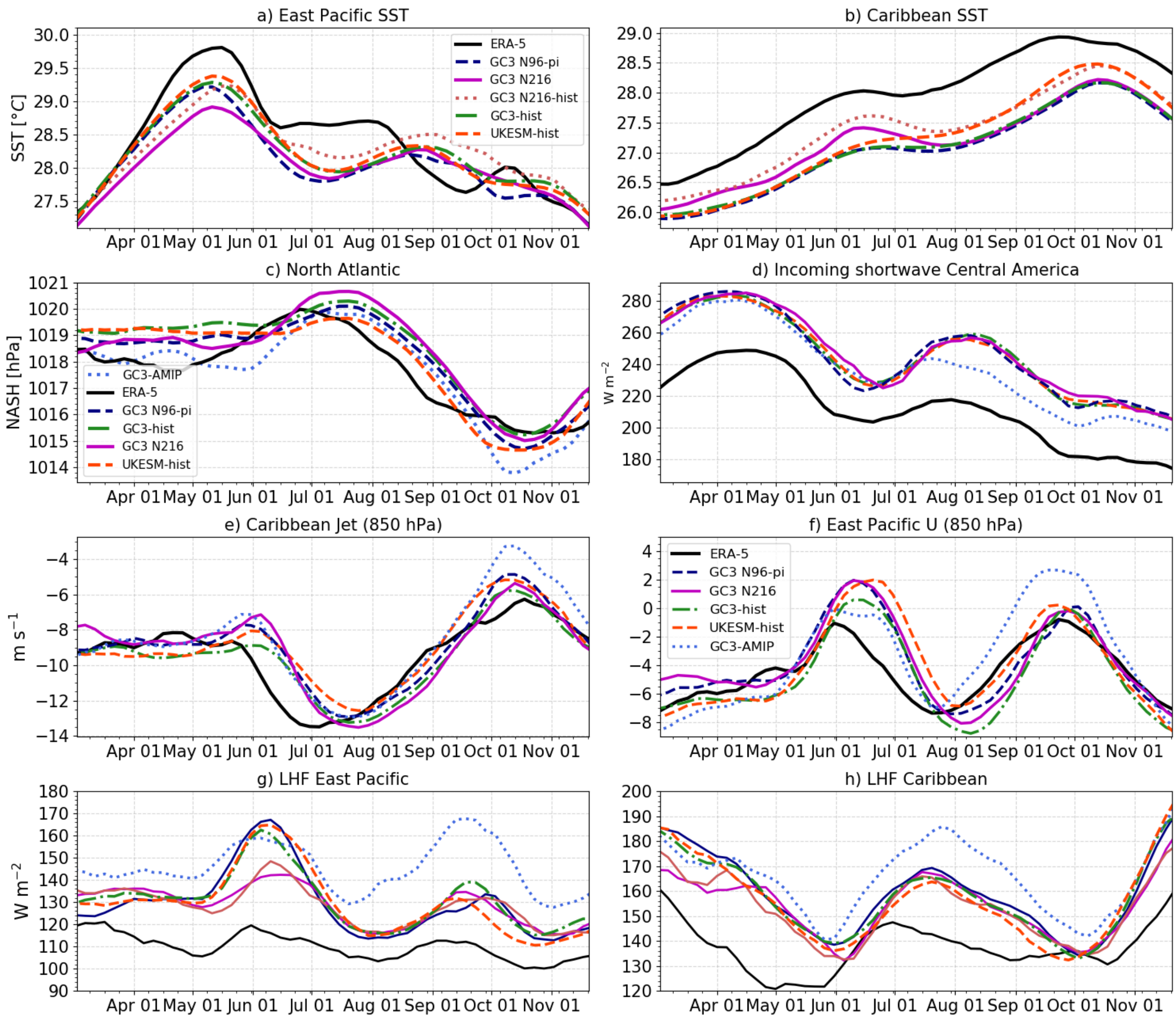

Figure 4.4: Pentad-mean seasonal cycle of indices associated with the MSD in Central America and the Caribbean.

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 mean precipitation rate.

Figure 4.3 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 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 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.

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 4.4. 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. 4.4g, 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 LHFs 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 the ERA5. This analysis suggests that the biases shown in Figure **??** are mostly coming from the period prior to the MSD. The models reasonably simulate the decrease in rainfall during the MSD (Figure 4.3) 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 4.1a.

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 4.5 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

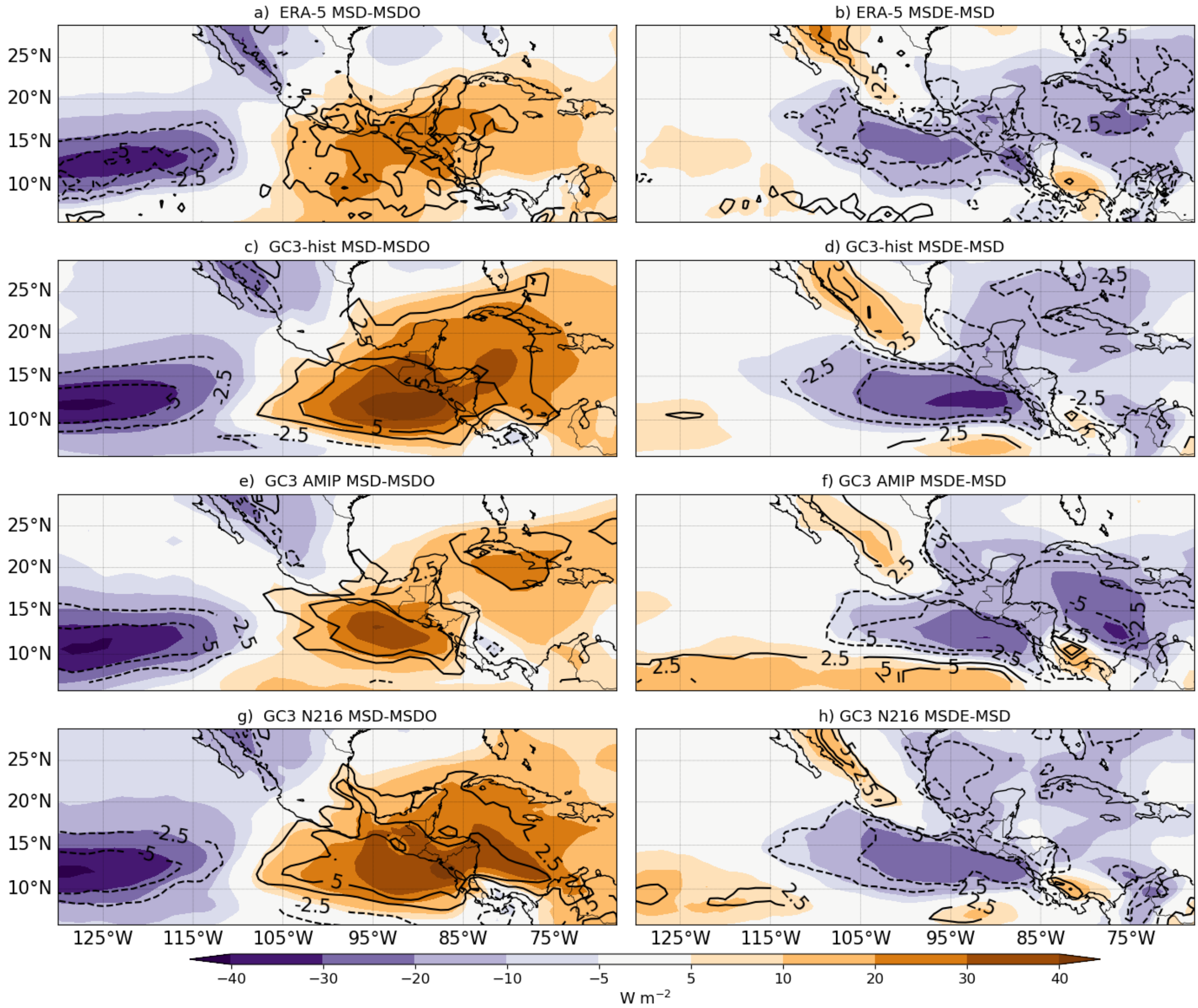

Figure 4.5: Out-going longwave radiation (OLR) [W m^{-2}] (shaded) and ω 500-hPa [10^{-2} Pa s^{-1}] (line contours) differences between the MSD and MSDO and the MSDE and MSD.

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 4.6 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 4.4f). During the MSD, the winds in the EP become modestly strong eaterlies associated the easterly flow from the Caribbean Sea that crosses over 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.

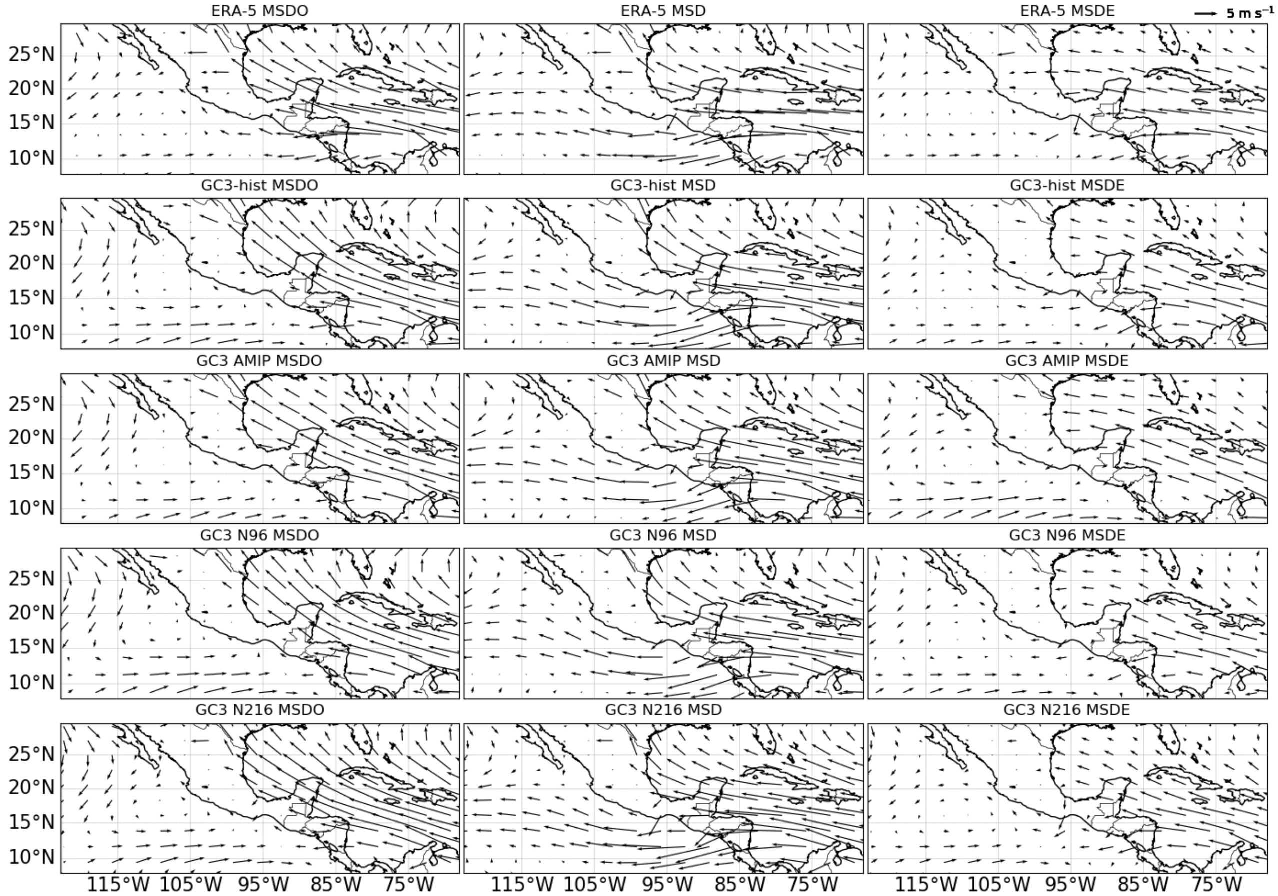

Figure 4.6: As in Figure 4.5 but showing wind vectors at the 850 hPa level.

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 4.4f, 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 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 2.3. Figure 4.7 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

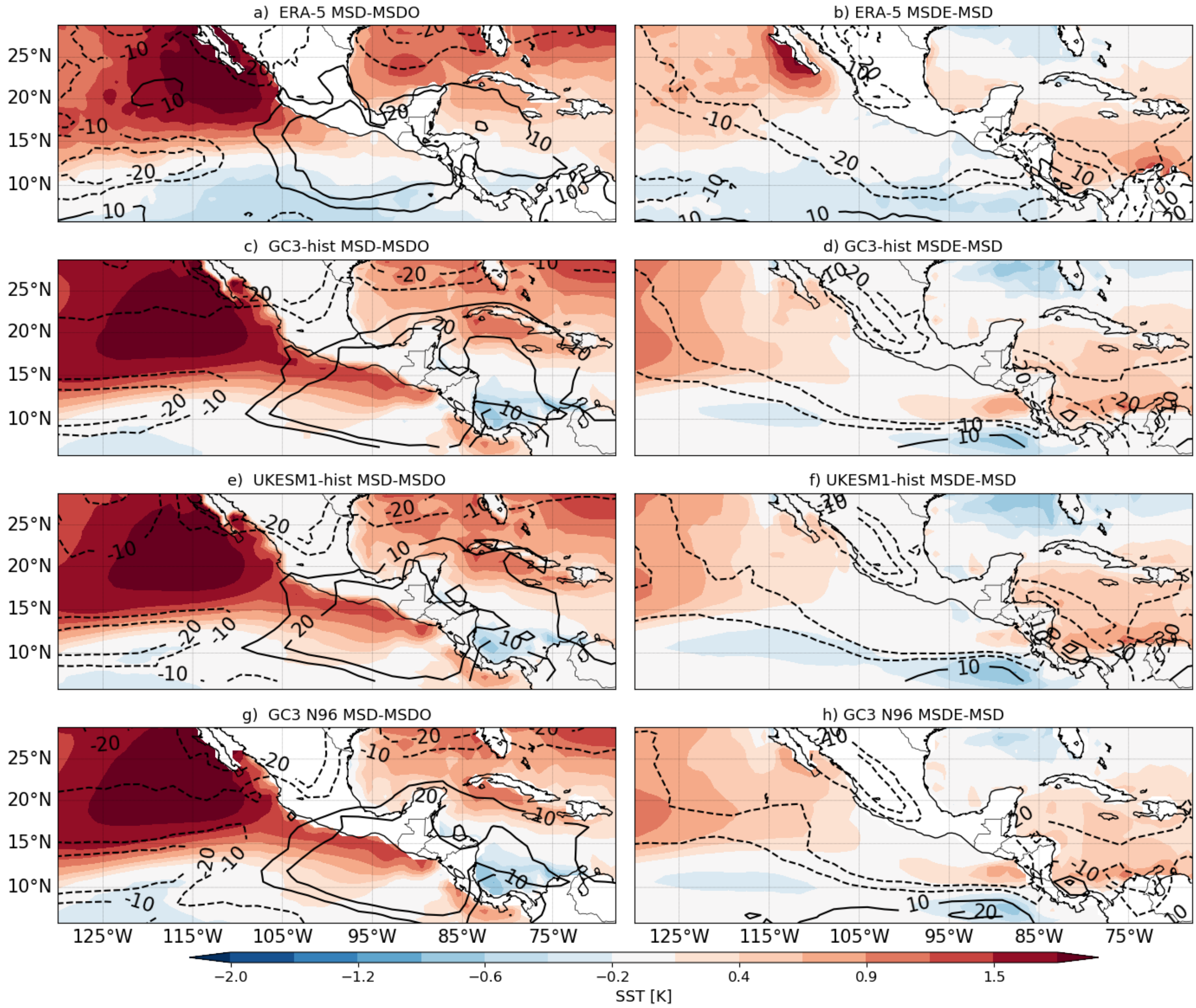

Figure 4.7: As in Figure 4.5 but the anomalies are shown for SSTs [K] (contours) and incoming shortwave radiation [W m−2] at the surface (line-contours). Incoming shortwave is defined such as negative differences imply less incoming shortave and positive anomalies represent more incoming shortwave at the surface.

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 4.4d. 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).

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

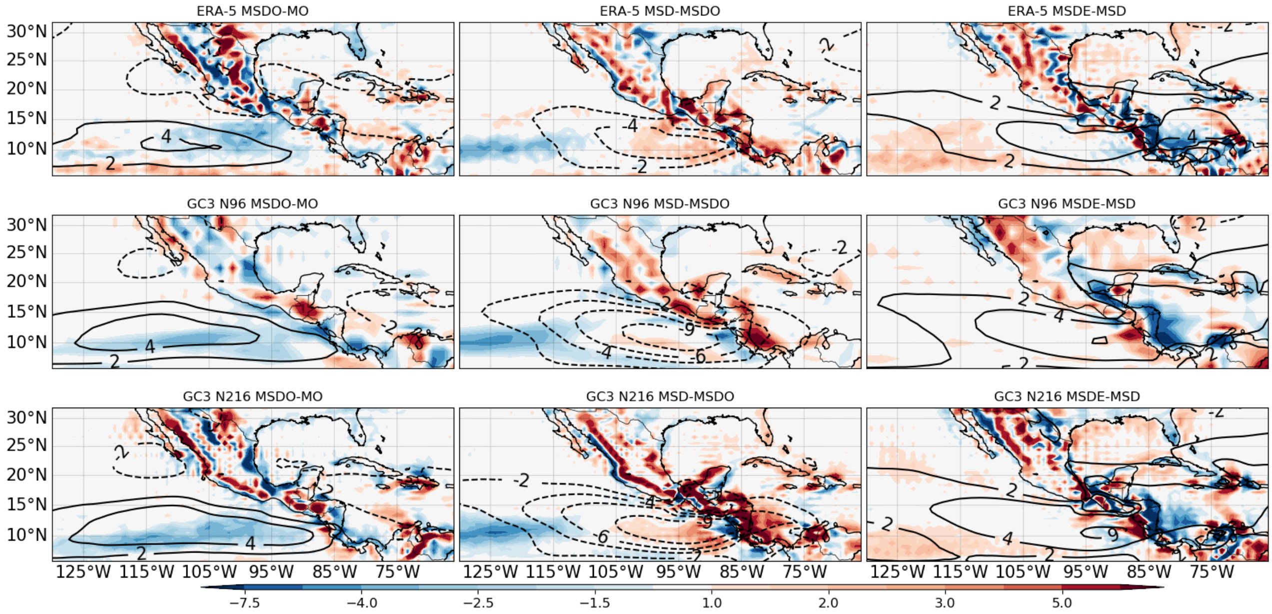

Figure 4.8: As in Figure 4.5 but showing in shading, moisture flux divergence $\nabla \cdot \vec{u}q$ at the 850 hPa level with units of 10^{-7} s^{-1} kg / kg and zonal wind anomalies (line contours) in m s^{-1}.

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 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 4.8 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. 4.6).

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. 4.4e), synchronized with the start of the MSD. The zonal wind anomalies in the MSD-MSDO panels in Figure 4.8 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.

4.1.2 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 2.3) 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.

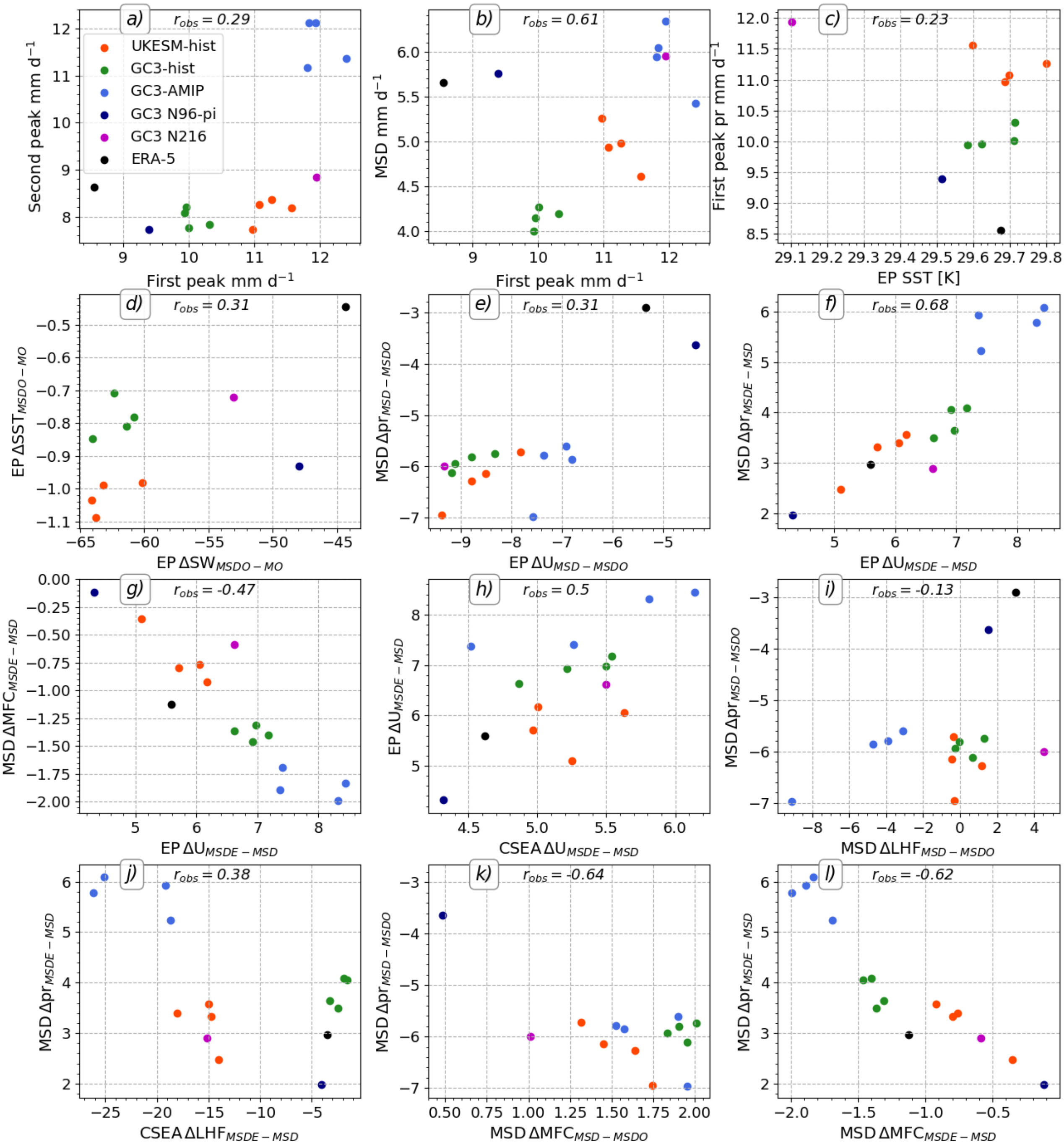

Figure 4.9: Scatter plot of the (a, b) area-averaged precipitation over land (Box in Figure **??**) during the different stages of the MSD. (c) scatter of the East Pacific SSTs against the precipitation over landduring the first peak period. (d-l) show the scatter differences in several variables between 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 [m s^{-1}], ΔMFC [10^{-11} s^{-1}], ΔSW and ΔLHF [W m^{-2}] and Δpr mm d^{-1}. The Pearson correlation coefficient for the 38 yr of reanalysis or observations (r_{obs}) is shown for each panel.

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 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 4.9a,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 4.9c 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 4.9d 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 4.4d), 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 4.9d, 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 4.6) 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. 4.4e). 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 4.9e, 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. 4.9g) 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. 4.9h). 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 4.4g, 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 4.9i,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 4.9k, 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.

Appendices

Cor animalium, fundamentum eft vitæ, princeps omnium, Microcofmi Sol, a quo omnis vegetatio dependet, vigor omnis & robur emanat.

The heart of animals is the foundation of their life, the sovereign of everything within them, the sun of their microcosm, that upon which all growth depends, from which all power proceeds.

— William Harvey **?**

A
Review of Cardiac Physiology and Electrophysiology

Contents

Appendices are just like chapters. Their sections and subsections get numbered and included in the table of contents; figures and equations and tables added up, etc. Lorem ipsum dolor sit amet, consectetur adipiscing elit. Sed et dui sem. Aliquam dictum et ante ut semper. Donec sollicitudin sed quam at aliquet. Sed maximus diam elementum justo auctor, eget volutpat elit eleifend. Curabitur hendrerit ligula in erat feugiat, at rutrum risus suscipit. Pellentesque habitant morbi tristique senectus et netus et malesuada fames ac turpis egestas. Integer risus nulla, facilisis eget lacinia a, pretium mattis metus. Vestibulum aliquam varius ligula nec consectetur. Maecenas ac ipsum odio. Cras ac elit consequat, eleifend ipsum sodales, euismod nunc. Nam vitae tempor enim, sit amet eleifend nisi. Etiam at erat vel neque consequat.

A.1 Anatomy

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A.2 Mechanical Cycle

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A.3 Electrical Cycle

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A.4 Cellular Electromechanical Coupling

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The first kind of intellectual and artistic personality belongs to the hedgehogs, the second to the foxes . . .

— Sir Isaiah Berlin **?**

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