Planetary Dynamics of the Western Atlantic Midsummer Drought and its Relationship to the Asian Monsoon

Patrick Kelly

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UNIVERSITY OF MIAMI

PLANETARY DYNAMICS OF THE WESTERN ATLANTIC MIDSUMMER DROUGHT AND ITS RELATIONSHIP TO THE ASIAN MONSOON

By

Patrick Kelly

A DISsertation

Submitted to the Faculty of the University of Miami in partial fulfillment of the requirements for the degree of Doctor of Philosophy

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PLANETARY DYNAMICS OF THE WESTERN ATLANTIC MIDSUMMER DROUGHT AND ITS RELATIONSHIP TO THE ASIAN MONSOON

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Warm season precipitation in the subtropical Americas is characterized by an abrupt drying in July and August known as the midsummer drought (MSD). The ultimate cause of the MSD is investigated in the context of large-scale atmospheric dynamics using observational data as well as a series of dedicated global modeling experiments. Observational analysis indicates that the climatological time signature as well as interannual variability of the MSD occurs in concert with the westward displacement of the North Atlantic subtropical high (NASH) and sign change of the mean barotropic zonal wind near 20-30ºN. Decomposition of a fully-closed zonal momentum budget reveals stationary eddy momentum flux divergence (SEMFD) on the western sector of the Asian monsoon Tibetan high as the primary mechanism governing the occurrence of zonal mean easterlies at subtropical latitudes in July. A stronger Asian monsoon in midsummer is significantly correlated with stronger subtropical easterlies and consequently, with a stronger MSD in the western Atlantic (WATL).

These results form the observational basis for a series of experiments using the Community Atmosphere Model version 4 (CAM4). The effects of a progressively enhanced Asian monsoon on the barotropic zonal wind are examined in CAM4 with an emphasis on downstream summer climate impacts in the greater WATL. The strength of
the monsoon is controlled by making the South Asian land surface darker (via lower soil albedo), leading to forcings of 5, 10, and 20 $\text{W m}^{-2}$ in net radiation. This leads to an enhanced Rossby wave source region over the Balkan Peninsula at 45°N, northwest of the upper level Tibetan high. Equatorward propagation of Rossby waves causes SEMFD to the south of this source region which produces easterly tendencies of the barotropic part of the mean zonal wind in the subtropics. As the easterly mean flow strengthens, so do low-level easterlies across the subtropical Atlantic, leading to a westward displacement of the NASH on its equatorward flank. The western intensification of the NASH causes drying in the WATL and neighboring land masses primarily due to near surface wind divergence in the anticyclone. Coupled air-sea interactions exhibit a positive enhancement of these monsoon-driven impacts, particularly off the Pacific coast of Mesoamerica, where increased easterly trades on the southwest corner of the NASH causes evaporative cooling of sea surface temperatures. These modeling results confirm the mechanistic hypothesis deduced from observations and are demonstrated to be insensitive to how heating over South Asia is introduced in the model. Related impacts of the mechanisms found here include steering of tropical cyclones on the western edge of the NASH, and interpretations for paleoclimate signatures.
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I am deeply grateful to my wife Maria for her unwavering support and confidence in me, and also for her recognition of the long hours away from home needed to complete this work. My two sons Dylan and Evan, keep me grounded and well-rounded as a person. They unconsciously re-inspire my love for learning about the natural world and show me new perspectives in life.

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CHAPTER 1: INTRODUCTION

1.1 The MSD

The annual march of the seasons is driven by smoothly varying changes in insolation as the sun crosses the equator leading to corresponding changes in surface heating. The seasonality of precipitation in the tropical Americas generally reflects these changes in radiative forcing. Deep, moist convection migrates meridionally with the sun, characteristic of monsoon systems over the Americas with a maximum and minimum in rainfall during the summer and winter, respectively (Vera et al., 2006). Over the greater subtropical western Atlantic¹ (WATL; box in Fig. 1) however, there are two identifiable peaks in precipitation during the boreal summer (hereafter, “summer”) rainy season. The bimodal precipitation distribution in the region is characterized by a July-August minimum in-between the late spring and late summer peaks (Fig. 1.1), and is an instance of a singularity (Wang and Xu, 1997; Mapes et al., 2005), or sub-seasonal kink in the annual cycle of precipitation.

Known colloquially as “Veranillo” or “Canicula” by local farmers for some time, this so-called midsummer drought (MSD) was first reported in the scientific literature by Portig (1961). Even though the rainfall minimum may have only moderate amplitude, the associated decrease in cloudiness under summer sun can also cause substantial increases

¹ Note that throughout this text, the term “western Atlantic (WATL)” also includes the extreme eastern Pacific off the coast of Central America as defined by the box in Figure 1.1
in temperature and evaporation, contributing to true hydrological drought conditions such as crop stress (e.g. Allen et al., 2010). The MSD is a natural sub-seasonal anomaly of the climate system not readily explained by variations in external solar forcing. *What internal non-linear dynamics of the climate system might then govern this high frequency anomaly?*

Figure 1.1: Seasonal precipitation anomaly for (a) JJA minus the annual mean and (b) DJF minus the annual mean. (c) Annual cycle of precipitation over western Atlantic (WATL) domain indicated by the black box. Data is from CMAP climatology averaged from 1979-2008.
Various mechanisms have been proposed to explain the MSD. Most notably, Magaña et al. (1999) hypothesized that the MSD could be explained by local air-sea feedbacks between sea surface temperature (SST), convection, and surface insolation off the Pacific coast of southern Mexico (southwest corner of box in Fig. 1) which drive fluctuations in the locations and intensity of the intertropical convergence zone (ITCZ). However, sub-seasonal SST anomalies in the region are fairly small in amplitude and spatial coverage (Small et al., 2007) and are thus unlikely to be the primary or ultimate cause of the MSD. The MSD is a large-scale phenomenon, encompassing the Caribbean, Gulf of Mexico, and neighboring landmasses (see Fig. 1.1) in addition to the coastal Pacific region identified by Magaña et al. (1999).

One clue to the cause of the MSD lies in the ability of many coarse atmosphere general circulation models (AGCM) to simulate it (Mapes et al., 2005; Rauscher et al., 2008), suggesting the physical mechanisms underpinning it are not too fine-scale or subtle. Figure 1.2 shows the character of midsummer precipitation in the WATL in both the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP, Xie and Arkin, 1997) climatology, as well as a super ensemble of 10 different AGCMs from the monsoon Atmospheric Model Intercomparison Project (AMIP; Kang et al., 2002). The total climatological time series of precipitation over the WATL is shown in the right panels for CMAP observations and the AMIP ensemble mean. To better isolate anomalies at sub-seasonal timescales, the annual (365-day period) and semiannual (183-day period) Fourier harmonics are subtracted from the total time series at each grid point to produce a high frequency (HF) time series. Left panels (Fig. 1.2 a,c) show maps of this HF time series averaged over July and August.
The large-scale MSD is reasonably well simulated in this 10-member ensemble. Midsummer dry anomalies are pervasive across the WATL domain in both observations and models, and the bimodal distribution of warm season precipitation is well captured. One distinction, however, is that the area of maximum drying (defined using HF time series) is located over southern Mexico in the AMIP-mean (Fig. 1.2c), as opposed to over the eastern Pacific in CMAP (Fig. 1.2a). This bias might stem from inadequate spatial resolution over Central America or perhaps in errors in the eastern Pacific ITCZ in uncoupled atmosphere models. Coupled processes, as suggested by Magaña et al. (1999), may be necessary to explain such subtle differences of the MSD (see Chapter 5.2). Nonetheless, the fact that several AGCMs simulate the large-scale midsummer drying across the WATL as in observations, points to a leading role of large-scale atmospheric dynamics in governing the MSD.

1.2 Mean Easterlies and the Westward Displaced NASH

Hasteranth (1967; 1976; 1978) was the first to suggest the role of the North Atlantic Subtropical High (NASH) as a possible cause of the MSD, which happens to strengthen and extend westward in July (see Figs. 1.3-1.4). More recently, several studies have also emphasized the importance of the NASH for the MSD (e.g. Mapes et al., 2005; Gamble et al., 2007; Gamble and Curtis, 2008; Kelly and Mapes, 2011) as envisaged by Hastenrath. The NASH can modulate precipitation in the WATL through a variety of processes. Anticyclonic vorticity anomalies associated with the NASH lead to changes in low-level divergence through Ekman pumping. The increase in subsidence that
accompanies an increase in SLP may also lead to changes in the vertical stratification of temperature and/or moisture, leading to changes in convective instability (Knaff, 1997). The NASH can affect precipitation through changes in water vapor fluxes since moisture transport patterns roughly follow SLP contours. Also, the midsummer enhancement of the NASH in the WATL strengthens the easterly trades or Caribbean Low-Level Jet (CLLJ), which is a key mode of moisture transport in the region (Wang et al., 2007; Muñoz et al., 2008). While the relative role of such process-level mechanisms of rainfall reduction in observations need further elaboration, the large-scale intensification of the NASH around July appears key to the onset of the MSD.

Figure 1.2: High frequency precipitation anomaly (annual and semiannual Fourier harmonics removed) centered on July and August for CMAP observations (a) and a 10-member model ensemble from AMIP (c). The total annual time series for the WATL region in CMAP (b) and AMIP-mean (d).

2 The physical mechanism relating the western intensification of the NASH to drying in the WATL is shown to be dominated by low-level wind divergence in CAM4 model experiments (Chapters 4.3 and 5.2).
While the intensification of the NASH is a proximate cause of the MSD, it begs the next question: *What causes the midsummer strengthening of the NASH?* Several studies (e.g. Rodwell and Hoskins, 2001; Seager et al., 2003; Miyasaka and Nakamura, 2005; Nigam and Chan, 2009) have examined the gross seasonality of the NASH and its Pacific counterpart— for instance, to explain why the subtropical anticyclones are strongest in summer, not winter. However, little research has been done at the refined level of examining sub-seasonal fluctuations of the NASH on the time scale of interest here (e.g. June to July changes). The low-level anticyclonic circulation of the NASH is primarily maintained by Sverdrup vorticity balance between the horizontal advection and divergence terms in the eastern Atlantic basin: Heating contrasts at the eastern boundary of the Atlantic drive equatorward flow and positive planetary vorticity advection. This *B*-effect is balanced by vortex tube shrinking (negative tendency) associated with subsidence and low-level divergence off the coast. Radiative cooling of low-level clouds further enhances this subsidence (Seager et al., 2003; Miyasaka and Nakamura, 2005), and remote dynamical influences— like Rossby waves emitted from monsoons— may also contribute (Rodwell and Hoskins, 1996; Rodwell and Hoskins, 2001). Westward dispersion of Rossby waves from this east basin vorticity source (Rossby wave source) then shapes the NASH in this view.

This eastern-basin driven view is consistent with the mean eastward location of the NASH, but the southwest corner of the NASH extends into the WATL around July, whose low-level easterly trade winds form the CLLJ. The July climatology of the NASH is depicted in Figure 1.3, which shows the 850 hPa geopotential height (*Z*850) and wind fields (Fig. 1.3a) along with the corresponding 850 hPa relative vorticity (Fig. 1.3b). The
protrusion of the NASH into the WATL in July results in a well-defined midsummer (July through August) minimum in the annual cycle of 850 hPa vorticity over the WATL (left box in Fig. 1.3b), with a precipitous negative anticyclonic tendency from June to July (Fig. 1.3c). On the other hand, the eastern Atlantic (EATL, right box in Fig. 1.3b) exhibits a corresponding maximum in vorticity in midsummer (Fig. 1.3c).

Figure 1.3: (a) July Z850 contours overlain with the vector wind field. The contour interval is 40 m. (b) 850 hPa relative vorticity in July. Right-hand and left-hand boxes represent the eastern Atlantic (EATL) and western Atlantic (WATL) domains, respectively (c) Time series of 850 hPa vorticity from the EATL (dashed line) and WATL (solid line) boxes. Daily data is smoothed by a 5-day box-car average and all data are from MERRA climatology.
Fourier harmonic time series analysis (as in Fig. 1.2) is performed to further illustrate the character of NASH variability in midsummer. Figure 1.4 shows maps of the filtered HF SLP field (in color) for early summer (May plus June; top) and midsummer (July plus August; bottom) in the National Center for Environmental Prediction (NCEP) Global Reanalysis v.1 (GR1; Kalnay et al., 1996). The total summertime SLP field (black contour) is also shown, indicating the mean eastward position of the NASH. The entire basin-scale NASH circulation doesn’t ubiquitously strengthen in midsummer. Rather, an east-west dipole structure is clearly evident on the equatorward flank of the NASH near 20-30°N, consistent with interpretation of Figure 1.3. In May-June, positive SLP (or “anticyclonic”) anomalies cover the eastern Atlantic with negative SLP (or “cyclonic”) anomalies over North America and the WATL. By July, this dipole reverses phase with negative anomalies in the eastern Atlantic and positive anomalies over the WATL. While SLP is examined here, identical analysis using other spatially smooth fields (e.g. geopotential height or stream function) yields qualitatively similar results. Other studies of the MSD have emphasized land-sea thermal contrasts (e.g. Mapes 2005; Small et al. 2007), which might lead to a local western intensification of the NASH due to a local Walker cell forced by continental heating over North America (Small et al, 2007). While not inconsistent with such a mechanism, the out-of-phase midsummer SLP changes across the subtropical Atlantic (Fig. 1.4) point towards a westward displacement of the entire NASH circulation, rather than just a local expansion on its western periphery. If so, it implicates larger than basin scale factors in governing the midsummer strengthening of the NASH in the WATL.
One clue to the cause of this westward displacement rests in the annual cycle of the zonal mean zonal wind \([u]\). The planetary zonal mean zonal wind turns easterly near 20-30\(^\circ\)N in midsummer, which could cause an east-west displacement of the NASH and therefore act as an ultimate cause of the MSD. Figure 1.5a shows a latitude-time climatology of the vertically and zonally averaged \(\langle[u]\rangle\), and a time series of \(\langle[u]\rangle\) averaged across 20-30\(^\circ\)N is also shown in Figure 1.5b. (Here \([\ ]\) denote a planetary

![Figure 1.4](image_url)

**Figure 1.4** High-frequency (annual and semi-annual harmonic removed) SLP in (a) May and June (b) July and August. Black contours in both panels represent the mean climatological SLP field in summer. Data are from NCEP reanalysis climatology.
zonal mean and $\langle \cdot \rangle$, is a density weighted vertical mean). Zonal mean easterlies protrude to 30°N in July and August. In Figure 1.5b., there are two critical points where the mean zonal flow in the subtropics changes sign: from westerly to easterly (beginning of July) and then back to westerly (middle of September). The timing of these changes correspond to the timing of 850 hPa vorticity changes across the Atlantic basin (Fig. 1.3c). Indeed the zonal wind at 850 hPa ($u_{850}$) across the Atlantic (red line in Fig. 1.5) also has an easterly maximum in July and closely resembles $\langle [u] \rangle$ in midsummer. Since $\langle [u] \rangle$ is a more tractable quantity, we focus on it hereafter. Study of the $\langle [u] \rangle$ budget, a global quantity, may lend considerable insight into the seasonal evolution of regional flow and climate impacts in the Atlantic Basin.

This work seeks a fundamental understanding of the MSD in the context of the large-scale dynamics of the climate system. Examination of basic climatology above lays the foundation for the argument that the MSD is ultimately governed by planetary scale dynamics, a hypothesis which will be further tested in interannual variations (Chapter 3) and model experiments (Chapter 4) below. An ultimate explanation of the MSD is offered, relating summer climate impacts in the WATL to the Asian monsoon through subtle, yet significant changes in the mean zonal wind $\langle [u] \rangle$. It is hypothesized that the western displacement of the NASH in midsummer, and hence drying in the WATL, is driven by the onset of negative (westward) advective tendencies by easterly $\langle [u] \rangle$. If this working hypothesis is correct, then we first need to understand the story of $\langle [u] \rangle$ in summer.
Figure 1.5: (a) Latitude-time plot of $\langle [u] \rangle_{\text{clim}}$. (b) Time series plot of $\langle [u] \rangle_{\text{clim}}$ averaged across 20-30°N (black line) and U850 averaged across 20-30°N in the Atlantic sector (red line). The zero line is dashed in both plots and data are from MERRA climatology.
CHAPTER 2: Mean Zonal Momentum in Summer

2.1 Closed MERRA $\langle [u] \rangle$ Budget

A momentum budget is deconstructed in this section in order to explicate the extension of zonal mean easterlies into the subtropics in midsummer (Fig. 1.5) Wind tendencies are available in the National Aeronautics and Space Administration’s (NASA) Modern-era Retrospective Analysis for Research and Applications (MERRA; Bosilovich et al., 2006) in addition to standard zonal and meridional wind fields, allowing for a complete decomposition of the zonal momentum budget. The total tendency is divided among 5 source terms so that the Eulerian rate of change of the zonal mean wind can be written as:

$$\frac{\partial \langle [u] \rangle}{\partial t} = \langle [DUDT_{dyn}] \rangle + \langle [DUDT_{GW}] \rangle + \langle [DUDT_{TKE}] \rangle + \langle [DUDT_{MST}] \rangle + \langle [DUDT_{ANA}] \rangle$$

(2.1)

The first term on the RHS in Equation 2.1 is the zonal wind tendency due to atmosphere dynamics, the sum of the pressure gradient force, Coriolis force, and advection terms. Note that mountain torques are implicit in this term (the residual of the zonal mean pressure gradient where mountains intrude into the atmosphere). Term I will be further broken down in Chapter 2.2.

Term II is due to the gravity wave drag scheme and Term III represents the turbulent transport of momentum (mostly in the planetary boundary layer), including
surface friction. Together, these two terms constitute physical drag of zonal momentum. The fourth term, includes all moist processes, chiefly vertical mixing of momentum by convection.

Finally, $DUDT_{ANA}$ is the “analysis tendency” of the zonal wind, an artificial forcing term introduced during the Incremental Analysis Update (IAU, Bloom et al., 1996) used in MERRA. $DUDT_{ANA}$ reflects the difference between a short-term forecast and the observed (analyzed) zonal wind field at a given time. Its climatology reflects systematic errors in the model.

Figure 2.1: Daily time series of the major terms in Equation 2.1 for vertical mean tendencies of the zonal wind in MERRA at 20-30°N. Daily data are from MERRA climatology and are smoothed by a 5-day box-car average.

Figure 2.1 shows an annual time series of the major terms in Equation 2.1, averaged across 20-30°N. The observed $\langle [u] \rangle$ tendency (LHS of Eq. 1) has a decreasing trend from winter to summer with maximum negative values in June which jump sharply
to positive values in the middle of July, implying a minimum in \( \langle [u] \rangle \) at this time (consistent with Fig. 1.5). \( \langle [DUDT_{DYN}] \rangle \) goes negative in late-May with a minimum around July, as large-scale dynamics exert westward force during the warm season. The annual cycle of \( \langle [DUDT_{TRB}] \rangle \) resembles \( \langle [u] \rangle \) in Fig. 1.5 but is of opposite sign since frictional drag opposes the direction of the flow. In the subtropics, surface drag acts as a momentum source in midsummer, providing positive zonal momentum from the solid earth to the atmosphere when the mean zonal flow goes negative in July and August (Fig. 1.5). \( \langle [DUDT_{ANA}] \rangle \) is substantially positive throughout the year, implying systematic model errors. These results are also summarized in Table 2.1, which lists the value of each term averaged over June and July.

The summation of all 5 tendencies is equal to the total rate of change of \( \langle [u] \rangle \) and is on the order of negative 3-4 m s\(^{-1}\) mo\(^{-1}\) yielding a net acceleration from 1 June–1 August of around negative 7 m s\(^{-1}\) (Fig. 1.5). Equation 2.1 represents a fully closed zonal momentum budget in MERRA as the net tendency from the RHS approximately equals the observed tendency from June to July. \( \langle [DUDT_{DYN}] \rangle \) is the largest term at approximately -5 m s\(^{-1}\) mo\(^{-1}\) and is the primary source for the negative \( \langle [u] \rangle \) tendency in summer. \( \langle [DUDT_{TRB}] \rangle \) is a negative source tendency for \( \langle [u] \rangle \) acting to reduce westerly momentum in June by friction. The tendencies from moist physics and gravity wave drag are negligible and can be ignored. \( \langle [DUDT_{ANA}] \rangle \) on the other hand is not negligible, and its positive value on the order of 3 m s\(^{-1}\) mo\(^{-1}\) acts to partially offset the negative tendency from \( \langle [DUDT_{DYN}] \rangle \). As \( DUDT_{ANA} \) represents unknown model biases, it unfortunately can’t be readily diagnosed. Fortunately, however, its climatological time signature does not account for the negative acceleration of \( \langle [u] \rangle \) in summer. Since \( \langle [DUDT_{DYN}] \rangle \) is the
only significant negative tendency, it is decomposed further in the following section to identify the role of eddies as the primary source of the negative mean zonal wind acceleration in summer.

<table>
<thead>
<tr>
<th>Term</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>DUDTDYN</td>
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</tr>
<tr>
<td>DUDTGWD</td>
<td>-0.05</td>
</tr>
<tr>
<td>DUDTTRB</td>
<td>-1.45</td>
</tr>
<tr>
<td>DUDTMST</td>
<td>-0.14</td>
</tr>
<tr>
<td>DUDTANA</td>
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<td>Sum</td>
<td>-3.40</td>
</tr>
<tr>
<td>LHS of Eq. 1</td>
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</tr>
</tbody>
</table>

Table 2.1: Individual zonal wind tendencies fields in MERRA for 20-30°N for June to July. Values are vertically and zonally averaged. The summation of all individual tendency terms as well as the actually tendency is also shown. Units are in m s\(^{-1}\) mo\(^{-1}\).
2.2 Dynamical Tendencies in Detail

The zonal mean of $DUDT_{DY}$ can be expanded as:

$$
\begin{align*}
[DUDT_{DY}] &= \left[ v \right] \left( f - \frac{1}{\cos \phi} \frac{\partial [u \cos \phi]}{\partial y} \right) - \left[ w \right] \frac{\partial [u]}{\partial p} - \frac{1}{\cos^2 \phi} \frac{\partial [u^* v^*]}{\partial y} \\
&- \frac{\partial [u^* w^*]}{\partial p} - \left[ \frac{\partial \Phi}{\partial x} \right] \\
&\text{(2.2)}
\end{align*}
$$

The notation here is standard with asterisks indicating a zonal eddy. Term I is the horizontal convergence of zonal momentum by the mean meridional circulation (MMC). This includes both the Coriolis deflection of the Hadley cell as well as advection of the mean zonal wind by the mean meridional wind. Term II is the vertical advection of zonal momentum by the MMC. Term III is eddy momentum flux convergence (EMFC). Terms IV and V are the vertical convergence of zonal momentum by the zonal eddies and the mountain torque term. MERRA’s pressure-level datasets omit grid points under ground so that the zonal mean pressure gradient can be computed on individual isobaric levels. As a check, the mountain torque term was calculated using surface pressure and elevation data following Lorenz and Hartman (2003), and the two techniques yielded nearly identical results.

The dominant terms in Equation 2.2 are EMFC and the horizontal convergence of momentum by the MMC (Terms I and III), so for brevity we show only show those in
Figure 2.2 for June and July (The values of all terms are listed in Table 2.2). At upper levels (~200 hPa) around 25°N, there is a peak in negative tendency from EMFC which is offset by positive tendency from the MMC. As eddies decelerate the wind aloft, the departure from geostrophic balance drives an MMC as air flows poleward down the meridional pressure gradient. The Coriolis force acting on this resultant MMC results in a compensating eastward acceleration near 25°N. Furthermore, mass continuity dictates that any mean meridional flow at upper levels is offset by an equal and opposite flow at low levels (neglecting net pole-equator mass shifts). This vertical compensation is identifiable in Figure 2.2b, as the MMC induces a negative acceleration at low levels underneath the positive acceleration aloft.

The mutual adjustment between eddies and the MMC is well known (e.g. Holton, 2004) and the strong cancellation between the two terms is clearly demonstrated in the Transformed Eularian Mean formulation of the zonal mean circulation (Andrews and McIntyre, 1976; Andrews et al., 1987). Ultimately, it’s the small imbalances between EMFC and the Coriolis torque on the MMC that force zonal mean flow changes at any one level. One of the difficulties in a budget analysis such as this one, however, is the ability to clearly identify the term responsible for the imbalance at any one level. This issue is mitigated by considering the vertically averaged form of Equation 2.2. Table 2.2 lists the column mean values of each of the 5 forcing terms from Equation 2.2 averaged across 20-30°N from June to July. EMFC stands out as the single dominant term, as all other terms are an order of magnitude smaller. Hence EMFC is responsible for column
mean forcing of the zonal mean zonal wind on the order of \(-7 \text{ ms}^{-1} \text{mo}^{-1}\) in the subtropics in midsummer. Moreover, the summation of all terms on the RHS of Equation 2.2 is nearly equivalent to the respective \(\langle [DUDT_{DYN}] \rangle\) value, indicating that the model’s actual dynamical tendency can be accurately decomposed into its individual source terms from daily resolution pressure-level outputs in MERRA. The vertically averaged form of Equation 2.2 can now be simplified as:

\[
\langle [DUDT_{DYN}] \rangle \approx -\frac{1}{\cos^2 \phi} \frac{\partial \langle [u^* v^*] \rangle}{\partial y} \tag{2.3}
\]

After substituting into our original tendency equation (Eq. 2.1), the MERRA \(\langle [u] \rangle\) budget can be simplified as:

\[
\frac{\partial \langle [u] \rangle}{\partial t} \approx \frac{1}{\cos^2 \phi} \frac{\partial \langle [u^* v^*] \rangle}{\partial y} + \langle [DUDT_{TRB}] \rangle + \langle [DUDT_{ANA}] \rangle \tag{2.4}
\]
<table>
<thead>
<tr>
<th>Term</th>
<th>Value</th>
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</tr>
<tr>
<td>Term 2</td>
<td>0.96</td>
</tr>
<tr>
<td>Term 3</td>
<td>-7.32</td>
</tr>
<tr>
<td>Term 4</td>
<td>0.58</td>
</tr>
<tr>
<td>Term 5</td>
<td>0.62</td>
</tr>
<tr>
<td>Sum</td>
<td>-4.55</td>
</tr>
<tr>
<td>DUDTDYN</td>
<td>-4.91</td>
</tr>
</tbody>
</table>

Table 2.2: Vertically and zonally averaged values of each term in Equation 2.2 at 20-30°N for June to July as well as for the actual model dynamical tendency. The summation of the five individual source terms is also given. Units are m s\(^{-1}\) mo\(^{-1}\)
Figure 2.2: (a) total EMFC and (b) horizontal convergence of zonal momentum by the MMC for June and July in MERRA climatology.
2.3 200 hPa EMFC Distribution

It is evident from above that the EMFC term is the cause for the negative acceleration of the vertically integrated \([u]\) in the northern summer subtropics. Since EMFC acts primarily at upper levels (Fig. 2.2), the eddy momentum torque at upper levels drives a tendency of the same sign at lower levels, via mass compensation in the MMC. This constraint simplifies our analysis greatly, allowing EMFC to be studied in more detail at upper levels.

Though mathematically, eddy momentum fluxes are only of formal interest to the \([u]\) budget in their zonal average form, analysis of the horizontal spatial distribution of \(u\ast v\ast\) sheds insight into the important regional processes responsible for this zonal mean forcing. Eddy momentum flux can also be further distilled down in time by making use of the simple temporal decomposition:

\[
\overline{u\ast v\ast} = \overline{u\ast v\ast} + \overline{u'v'}. \tag{2.5}
\]

Here, the total eddy flux for a given month is the product of 3-hourly \(u\) and \(v\) wind fields, the stationary eddy flux is the product of monthly \(u\) and \(v\) wind fields, and the transient eddy flux is simply the arithmetic difference of the two.

Momentum fluxes \(u\ast v\ast\) at 200 hPa averaged over June and July and overlain with the total wind field are shown in Figures 2.3 for transient eddies. A map of transient EMFC (TEMFC) is also shown in Figure 2.3 along with the latitudinal average calculated
across 20-30°N. The southwest-northeast phase tilt of short-wave trough and ridges—due to the typical meridional shear gradient (westerlies increasing with latitude)—causes zonal momentum to be transported poleward by transient disturbances. The transient eddy momentum flux (Fig. 2.3) is thus positive at all longitudes north of 30°N (implying a negative acceleration south of 30°N) as transient eddies redistribute zonal momentum as they move downstream in the prevailing westerlies. The TEMFC is rather small in any one location but the net contribution of transients to the zonal mean TEMFC at 20-30°N is not negligible. However, transient eddies peak in winter, not summer, so their time signature fails to explain the midsummer protrusion of tropical easterlies into the subtropics. Also, the geographical distribution of total (stationary and transient) EMFC in summer is shaped by stationary eddies forced by longitudinal heating gradients characteristic of a monsoon structure in the Northern Hemisphere (Chen, 2003; Chen, 2010).

In Northern Hemisphere summer, the south Asian monsoon is particularly dominant with far-reaching impacts and teleconnections across the globe (e.g., Lau and Peng, 1992; Lau and Weng, 2002; Ding and Wang, 2005). Diabatic heating associated with monsoon convection over southeast Asia forces a planetary-scale high pressure system in the upper troposphere (~200 hPa) known as the Tibetan High (TH). The TH is seen in Figure. 2.4, which shows observed geopotential height and winds at 200 hPa and precipitation for June-July. The TH is such a prominent anticyclonic flow that it makes closed contours of the total geopotential height (or stream function) field (Fig. 2.4). In longitude, the TH stretches all the way from the Greenwich Meridian on its western edge
to around $150^\circ$E to the east. The displacement of the upper level TH anticyclone to the west of convection is consistent with Gill’s (1980) theory of the westward spreading of a long Rossby wave pattern due to an off-equatorial heat source.

Figure 2.3: (a) Contoured 200 hPa transient eddy momentum fluxes ($u^*v^*$) in June and July overlain with total vector winds and (b) TEMFC ($-\partial u^*/\partial y$) of the eddy momentum fluxes in (a). (c) The longitudinal distribution of TEMFC in (b) calculated across 20-30$^\circ$N along with the mean value in each sector. Data are from MERRA climatology.
Figure 2.4 June-July climatology of 200 hPa geopotential height (red contours), horizontal winds (vectors) and precipitation (green-yellow contours). Circulation data are from MERRA (1979-2010) and precipitation data are from CMAP (1979-2010) data sets.

One implication of the TH for the global circulation is through its crucial role in the mean zonal momentum budget in the northern subtropics. Figure 2.5 is identical to Figure 2.3, except for stationary eddies instead of transient. The primary contribution to the zonal mean stationary EMFC (SEMFC) is from around 25°E longitude (Fig. 2.5b), which corresponds to the western edge of the TH (Fig. 2.5a). On its northwest corner (~35°N), the southwesterly flow around the TH interacts strongly with the oncoming westerlies, causing an area of enhanced positive momentum flux over the eastern Mediterranean. To the south (~20°N), there is negative eddy momentum flux from southeasterlies on the southwest corner of the TH. Hence, near 25°N, there is large SEMFC in the meridional direction as southwesterlies pump positive zonal momentum to the north and southeasterlies pump negative zonal momentum to the south. While similar reasoning of the opposite sense holds for SEMFC on the eastern flank of the TH (~100°E), the negative tendency from eddies near 25°E is only partially offset by the positive tendency to the east, due to the anomalously positive eddy momentum flux over the Mediterranean.
There is also a negative—albeit smaller in magnitude—zonal wind tendency due to SEMFC from the Tropical Upper Tropospheric Troughs (TUTTs) in the eastern Atlantic and Pacific basins (Fig. 2.5). TUTTs are cold-core lows that extend across the subtropical regions of the oceanic basins (Whitfield and Lyons, 1992). The positive (southwest-northeast) tilt of the TUTTs result in positive eddy momentum fluxes, with northeasterlies upstream of the trough axis and southwesterlies downstream. Situated above the subtropical anticyclones in summer (White, 1982), TUTTs are thought to be maintained by land-sea contrasts in heating and precipitation leading to cooling and subsidence over the eastern ocean basins. The generation of subsidence that leads to the formation of the stationary TUTT feature may originate from deep monsoon heating over the continents (Rodwell and Hoskins, 2001; Chen et al., 2001; Liu et al., 2001) or from shallow heating cooling couplets across the west coast of North Africa and North America (Miyasaka and Nakamura, 2005). The presence of TUTTs situated above the subtropical highs are no accident, but rather both are forced by deep subsidence driven by east basin cooling, with the positive vorticity source aloft balanced primarily by negative zonal advection (Miyasaka and Nakamura, 2005). At upper levels, the juxtaposition of the warm core Asian monsoon anticyclone (TH) and the cold core Atlantic cyclone (TUTT) to the west (see Fig. 2.4) appears to be especially important to the $[u]$ budget in the subtropics, causing the large positive eddy momentum flux over the Mediterranean (Fig. 2.5b). This regional contribution from Eurasia to the zonal mean 200 hPa SEMFC is greater than the Pacific and Atlantic Oceans by a factor of at least 3 (Fig. 2.5c).

The particular importance of monsoon driven SEMFC to zonal mean easterlies is also seen in Figure 2.6, which shows an annual time series SEMFC in the Eurasia sector.
(0-120°E) along with zonal mean of total (stationary + transient) EMFC at 30°N. The total negative acceleration from eddies from June to the middle of July is largely described by the SEMFC in Eurasia, which in turn is due to the western margin of the TH (Fig. 2.4). And since \([u]\) and \(<[u]\>\) go negative from June to July (Fig. 1.5), the climatological midsummer easterlies near 30°N appear to be primarily forced by the Asian monsoon. Given this strong dependence, we examine the interannual relationship between the Asian monsoon and SEMFC below, with implications on \(<[u]\>\) and the westward displacement of the NASH.

Figure 2.5: As in Figure 2.3, but for stationary eddies
Figure 2.6: Daily time series of the total (transient and stationary) EMFC at 30ºN (black line) and monthly time series of the contribution from SEMFC between longitudes 0-120ºE (red line).
CHAPTER 3: Interannual Connections

3.1 Asian Monsoon and SEMFC

The timing and intensity of the MSD in the WATL may be dynamically linked to the Asian monsoon through feedbacks on the zonal momentum budget in the subtropics. I suggest the interannual variability of the Asian monsoon and in particular, the TH, will drive the negative zonal mean eddy torque at 200 hPa (and therefore $\langle [u] \rangle$, see Chapter 2), an idea based on the climatological importance of SEMFC on the southwest corner of the TH (SWTH) in driving the zonal mean (Fig. 2.5).

Figure 3.1: Map depicting the various important regions and definitions used in this section upon which interannual indices and correlations are based. ‘IMI’ is the Indian monsoon index defined by Wang and Fan (1999) and ‘SWTH’ refers to the southwest corner of the Tibetan high where the maximum in SEMFD lies.
To test this idea, we employ the summer Indian Monsoon Index (IMI) defined by Wang et al. (2001). The IMI is calculated as the 850 hPa zonal wind (U850) difference between box 1 and box 2 in Figure 3.1. The IMI is a measure of the intensity of low-level inflow around and into the Indian monsoon trough, so a large and positive value indicates a “strong” year. Using MERRA data from 1979-2006, the mean June plus July IMI for each year is calculated. (The onset of the monsoon occurs sometime in June as is maturely established by July.) Figure 3.2 shows upper-lower tercile differences of strong minus weak IMI (9-year averages for each tercile) for eddy geopotential height at 850 hPa and 200 hPa, overlain with the total vector wind of wet years. During strong IMI years, there are anomalously low 850 hPa heights and cyclonic flow over and to the west of the enhanced convective activity over India, the Arabian Peninsula, and extending into the Mediterranean, consistent with the heating-driven dynamics discussed in Gill (1980) and Rodwell and Hoskins (1996). At upper levels, the TH is enhanced west of 80°E, with maximum positive anomalies near the eastern Mediterranean Sea. This finding is in agreement with previous diagnostic studies (Chaudhari et al., 2008; Rajeevan, 1993) showing 200 hPa anticyclonic anomalies associated with excessively warm and wet Indian monsoon years. Interestingly, the broad-scale anticyclonic circulation of the TH is not enhanced, with positive 200 hPa height anomalies only found west of India during strong years. In fact, the TH and the monsoon trough at 850 hPa over south China are diminished during strong IMI years– supporting the proposition that independent circulation indices are needed to describe the broader Asian monsoon (Wang et al., 2001).
Figure 3.2: June and July eddy geopotential height and vector wind differences of good minus bad IMI years. Black box indicates the SWTH region shown in Figure 3.1. Data are from MERRA climatology.
Figure 3.3: Composite differences of the June and July stationary eddy momentum flux ($u^*v^*$) for high minus low IMI years (color). Black lines denote the background climatological mean value with solid and dashed contours representing positive and negative values respectively. Data are from MERRA.
The enhanced anticyclonic flow centered over the Mediterranean during strong IMI years causes anomalous 200 hPa SEMFC \((-\partial uv/\partial y)\) on the extreme southwest corner of the TH as strong southeasterlies (negative \(u^*v^*\)) turn to weak southwestlies (positive \(u^*v^*\)) between 20-30°N in the box shown (Fig. 3.2a). This SWTH region corresponds to the western edge of the climatological SEMFC maximum in Figure 2.5, indicating a westward displacement of this SEMFC maximum during strong IMI years. The westward displacement of the climatological 20-30°N SEMFC maximum during strong IMI years can also been seen in Figure 3.3, which shows \(u^*v^*\) anomalies overlain with the climatological value. Guided by these composite-mean circulation differences, Figure 3.4 shows interannual scatter plots of the SWTH SEMFC (box in Fig. 3.2) versus the IMI in MERRA and NCEP-GR1 for 1979-2006. All values are averaged over June and July.

A stronger Indian summer monsoon significantly correlates with (and presumably is the cause of) a larger negative 200 hPa SEMFC (Figs. 3.4 a,b), which is the primary term governing \([u]\) tendency and the protrusion of easterlies into the subtropics in midsummer (Chapter 2). In other words, years with a strong cyclonic flow at low-levels over India have an enhanced upper-level anticyclone to the west, causing anomalous SEMFC as westerly (positive) \(u\) momentum is pumped northwards across the 30°N latitude line. Moreover, local values of SWTH SEMFC are positively correlated to the respective zonal mean value (Figs. 3.4 c,d), indicating that the SWTH is an important region governing interannual variability— in addition to the climatological mean (Fig. 6b)— of the 200 hPa zonal mean eddy torque in the subtropics.

Individual years are also color coded in Figure 3.4 based on the simultaneous (June and July) Niño 3.4 anomaly (Trenberth, 1997), with warm (red), cold (blue), and
neutral (black) color coding based on the respective tercile category. No clear relationships to the phase of ENSO are seen, indicating that the relationships here are not just aspects of ENSO variability. The interaction between ENSO and the Indian monsoon has a rich research history (e.g. Walker 1924; Bjerknes 1969; Pant and Parthasarathy, 1981). The negative relationship between warm ENSO events and monsoon precipitation (Yasunari, 1990; Webstar and Yang, 1992; Kirtman and Shukla, 2000) varies with time, and in particular, has diminished in recent decades, possibly due to global warming (Kumar et al., 1999; Ashrit et al., 2001). These findings may explain the lack of a relationship in Figure 3.4, since our relatively small sample of years start at 1979.

In comparing the two different data sets, MERRA has slightly higher correlation coefficients than NCEP-GR1 and there is some considerable disagreement in the details of individual values of SEMFC. The degree to which these reflect robust circulation biases or are an artifact of different horizontal resolution should be investigated further. Fortunately, interannual differences are qualitatively similar amongst both reanalyses and all correlations are statistically significant at the 95% confidence level.

3.2 $\langle [u] \rangle$ Based Composites

To test our underlying hypothesis outlined in the Introduction— that midsummer zonal mean easterlies in the subtropics cause westward advection of the NASH with implications for midsummer precipitation changes— we focus on interannual comparisons

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3 Also see Appendix C which demonstrates similar relationships in CAM4 forced by climatological SSTs.
of July minus June (JMJ) differences. JMJ differences are a useful metric, since this time signature captures the rainfall reduction (Figs. 1.1, 1.2), the corresponding strengthening of the NASH (Figs. 1.3, 1.4), and the negative acceleration of \(|u|\) (Fig. 1.5) in the subtropics (20-30°N). Interannual differences of \(|u|\) at 20-30°N for JMJ are indexed and tercile composites of individual fields are constructed based on this index. The upper tercile in this context refers to the 9-years (from 1979-2006) for when JMJ \(|u|\) is largest and most negative.

Figure 3.4: (a,b) Interannual scatter plots of June to July IMI versus the SEMFC on the southwest corner of the Tibetan High (SWTH; see figure 3.1). (c,d) Interannual scatter plots of SEMFC at the SWTH versus the planetary zonal mean at the same latitude band. Left-hand panels are for MERRA; Right-hand for NCEP reanalysis. Years are color-coded by ENSO 3.4 anomaly.
Upper minus lower tercile composites for precipitation, SLP, and 200 hPa SEMFC based on this JMJ $\{[u]\}$ index are shown in Figure 3.5. The zonal wind at 850 hPa (U850) is also shown (Fig. 3.5b), to confirm that it tracks $\{[u]\}$ as in climatology (see Fig. 1.5). Precipitation data used in Fig. 3.5 are from CMAP and wind and SLP data are from NCEP-GR1 (MERRA yields similar results). Data are first coarsened from its native 2.5° horizontal resolution to 10°, since the interest here is in \textit{large-scale} differences in magnitude between low and high years, where the statistical significance preferentially resides. For U850, the spatial averaging is increased to 60° in longitude, which is roughly the zonal scale of the Atlantic. This coarsening isolates the importance of basin-scale zonal wind changes in the Atlantic, which are the basis of our hypothesis. (At small scales, the wind changes are dominated by pressure differences that can be inferred from Fig. 3.5c).

Only differences for statistically significant grid points (at the 95% confidence level based on a \textit{t}-test) are plotted in Figure 3.5. Figure 3.5a shows that there is enhanced negative SEMFC over the SWTH (box in Figs. 3.1, 3.2) when JMJ $\{[u]\}$ change in the subtropics is most negative. The SWTH region corresponds to the climatological maximum in negative EMFC tendency identified previously, and is the primary source of the negative $\{[u]\}$ acceleration in midsummer, forcing both the mean climatology (Fig. 2.5) as well as its interannual variability (Figs. 3.3, 3.4). These composite differences of SEMFC are insensitive to the whether $\{[u]\}$ (Fig. 3.5a) or the IMI (Fig. 3.4a) is used as the base index, indicating that interannual variability of the Indian monsoon, SEMFC on the SWTH, and the negative acceleration of $\{[u]\}$, are robustly connected processes. The composite JMJ difference of the low-level zonal wind (U850) is shown in Figure 3.5b.
Figure 3.5: Composite differences of various fields for good minus bad $\langle [u] \rangle$ index years. Fields are for July minus June differences and only grid points significant using at the 95% level using a student $t$-test. SLP and wind data are from NCEP-GR1 and precipitation data are from CMAP. U850 is coarsened to 60° to emphasized basin scale changes. All other data are coarsened to 10°.

When the vertical mean $\langle [u] \rangle$ decreases from June to July in the subtropics, so does U850, with significant negative anomalies stretching westward across the Atlantic in the 20-30°N latitude band. But in the mid-latitudes around 45°N, JMJ U850 anomalies...
are positive over the Atlantic, implying a large-scale enhancement of the anticyclonic circulation (consistent with Fig. 3.5c). There is also a region of positive U850 anomalies southwest of the Indian subcontinent extending westward into Africa. This region around 10-20°N may reflect enhanced southwesterlies during years when the monsoon is more active.

SLP differences (Fig. 3.5c) show that the entire NASH (indeed, the zonal mean) is strengthened when $\langle [u] \rangle$ is most negative, with large positive anomalies over the WATL. July SLP falls over the Arabian Peninsula are consistent with the monsoon precipitation (heating) anomalies to the east—the Indian monsoon is wetter, with increased precipitation in July over India and Pakistan (Fig. 3.5d). Figure 3.5d also demonstrates that when there is a large and negative $\langle [u] \rangle$ JMJ change, the corresponding change in precipitation over the WATL is significant and negative.

### 3.3 Asian Monsoon and Western Atlantic Rainfall

As a consistency check of our key finding in Sec. 3.2 that a wetter Indian monsoon in July corresponds to a drier WATL (Fig. 3.5d), we examine JMJ precipitation composites (upper minus lower tercile) based on all of the important indices outlined above. Namely, JMJ differences of $\langle [u] \rangle$, SLP in the WATL, precipitation in the WATL, and the IMI are used as the base indices for targeting precipitation differences in Figure 3.6. Precipitation data is only coarsened to 5° here in order to capture the MSD over Central America, which due to its limited spatial extent, is noticeable absent in Fig. 3.5d with data at 10° resolution. The findings across all indices (four panels in Fig. 3.6) are
strikingly similar—there are wet anomalies over South Asia coincident with dry anomalies in the WATL. The magnitude of the difference between South Asia and the WATL also remains steady across the different base indices. The MSD onset occurs in July, but midsummer dry anomalies persist through August (Figs 1.1, 1.2). So, composite differences for precipitation for July plus August minus June are constructed in a similar fashion and shown in Figure 3.7. The same pattern emerges: Wet anomalies over South Asia are associated with dry anomalies in the WATL subtropics, particularly near Florida and Central America.

The robustness of this relationship is further confirmed in Fig. 3.8, which expands the precipitation compositing technique to other indices of monsoon intensity as well as observational data sets. The Webster-Yang (WY) monsoon index (Webster and Yang, 1992) is a measure of the large-scale vertical shear between the upper level easterlies and near surface westerlies over South Asia. The All Indian Rainfall (AIR) index (Parthasarathy et al., 1995), as its name implies, is a measure of precipitation over India (land only). JMJ difference of precipitation from CMAP as well as the Global Precipitation Climatology Project (GPCP; Adler et al., 2003) are composited on the IMI, WY, AIR indices in Figure 3.7. For clarity across multiple panels, precipitation data in Figure 3.7 is coarsened to $10^\circ$ to reduce noise, mindful that the small (in spatial extent) but robust drying signal over Central America is aliased at this resolution (see Figs. 3.5, 3.6). Different indices and data sets tell a consistent story. The consistency of this inverse relationship suggests that the Asian monsoon and the MSD are robustly related features whose interannual variability appear to be linked through the midsummer enhancement of subtropical easterlies and the intensification of the NASH. While all three monsoon
indices show wet anomalies in South Asia coincident with dry anomalies in the WATL, the IMI index shows the clearest signal with greater inter-hemisphere precipitation differences. The two major convective centers of the broader Asian monsoon–convection near India and that over the West Pacific–are not necessarily correlated on interannual timescales [Wang and Fan, 1999; Wang et al., 2001], The stronger signal in the top panels of Figure 3. may be due the fact that the IMI captures circulation and heating anomalies on the western margin of the monsoon near India, and it’s the western edge of the TH that is important in terms of westward Rossby wave dispersion and SEMFC (Fig. 2.5).

Given the high degree of serial correlation and spatial interdependence often present in meteorological fields, Livezey and Chen (1983) suggest that field (global) significance testing is required to ascertain statistical significance. To address this, we performed a series of Monte Carlo experiments (following Wolter et al., 1999) of the precipitation differences presented in Figures 3.6-3.8. 10,000 random time series each 28 years long (corresponding to the 1979-2006 base period) were created using a random number generator. These time series were subsequently sorted into upper and lower terciles. The JMJ precipitation data was then composited based on these indices from the randomly generated time series. Precipitation differences of upper minus lower tercile were then retested for local significance at each grid point. Finally, the number of significant grid points at the 95% confidence level (using the $t$-test) from the 10,000 experiments were arranged in a decreasing sequence with the 500th grid points marking the field significance threshold ($\alpha = 0.05$). For each base-index in Figures 3.6-3.8, the
number of significant grid points is greater than the relevant Monte Carlo criterion, indicating that the above composite differences of precipitation are meaningful.

Figure 3.6: July minus June precipitation composite differences based on various indices. Only grid points at the 95% confidence level (using a t-test) are shown. Precipitation data are from CMAP and SLP and wind data are from NCEP-GR1. Data are coarsened to 5deg resolution. See Figure 3.1 for definition of indices.
Figure 3.7: As in Fig. 3.6, but for July plus August minus June differences
Figure 3.8: July minus June composite precipitation differences composited on different Asian monsoon indices in (a) CMAP and (b) GPCP data sets. Data are coarsened to 10º resolution and only grid points significant at the 95% confidence level using a t-test are shown.
3.4 Correlation Analysis

Interannual time series of several key indices are examined here to extend the conclusions drawn from the previous section. There is a significant difference in JMJ SLP and precipitation over the WATL region dependent on the strength of the zonal mean easterlies in the subtropics (Fig. 3.5). Individual time series for JMJ SLP and JMJ precipitation in the WATL are constructed and compared to the respective $\langle [u] \rangle$ time series. Figure 3.9 shows scatter plots of JMJ differences for each individual year from 1979-2006.

There is a significant negative correlation ($r = -0.8$) between SLP and precipitation (Figs. 3.9 a,b), perhaps unsurprisingly as the midsummer strengthening of the NASH has previously been cited as the proximate reason for the rainfall deficit in the WATL (see Introduction). JMJ $\langle [u] \rangle$ also significantly correlates with SLP: years with strong easterlies also have a large and positive SLP rise in July (Figs. 3.8c,d). Furthermore, interannual variability of $\langle [u] \rangle$ also correlates significantly with precipitation in the WATL (Figs. 3.9 e,f). It appears that the strength of the midsummer easterlies, as viewed by JMJ $\langle [u] \rangle$ changes, is a key indicator of interannual variations of the NASH and hence, July drying in the WATL. Reassuringly, the same analysis for MERRA and NCEP-GR1 give consistent results (left and right panels in Fig. 3.9) and as mentioned previously, these year-to-year variations of sub-seasonal anomalies (JMJ differences) are not significantly correlated with the Niño 3.4 anomaly (color coding in Fig. 3.9).
Figure 3.9: (a,b) Interannual scatter plots of SLP versus precipitation in the WATL domain. (c,d) Scatter plots WATL SLP versus \( \langle |u| \rangle \) at 20-30°N (c,d) as well as WATL precipitation versus \( \langle |u| \rangle \) (e,f). Left-hand panels are for MERRA; Right-hand for NCEP-GR1. Values are for July minus June differences and years are color-coded by ENSO 3.4 anomaly.
Figure 3.10: Interannual time series of the key indices discussed in this text for the total time series (a) and low-pass filtered time series to remove frequencies with periods less than 7 years (b). All values are for July minus June differences. Precipitation is in units of mm d$^{-1}$, $\langle [u] \rangle$ is in m s$^{-1}$ and IMI is a normalized quantity. See Fig. 3.1 for definition of indices.

The correlations presented in Fig. 3.9 and the composite differences in Figs. 3.5-3.8 point to a significant relationship between the Indian monsoon, $\langle [u] \rangle$, and WATL precipitation on an interannual time scale. Figure 3.10 shows time series of these 3 key indices for JMJ to better understand the frequency of the interannual variability driving these correlations. NCEP-GR1 $\langle [u] \rangle$ data and IMI index are shown in Fig. 3.10 because of the longer record length (beginning in 1948), though results using MERRA are similar for the same period (post-1979). Fig. 3.10a demonstrates that in years when $\langle [u] \rangle$ JMJ is
large and negative, JMJ precipitation in WATL also tends to be large and negative while the IMI tends to be strong and positive (equivalent to stronger monsoon in July). These interrelationships appear to arise largely from year-to-year variability. In other words, they are not solely a result of a particularly strong correlation from one or two years (e.g. the 1982-83 and 1997-98 El Niños), or of a trend over the entire record length. A truly physical relationship should apply an all timescales (also see Appendix D). There is some indication of decadal variability, however, which is better seen by low-pass filtering the total time series shown in Fig. 3.10a. Figure 3.10b shows time series of the same indices but with a low-pass digital filtered applied to isolate periods longer than 7-years.

### 3.5 Summary

This diagnostic study revealed several key connections lending support to our hypothesis that are summarized in the schematic in Figure 3.11. Heating from precipitation related to the Asian monsoon is a primary processes governing eddy momentum fluxes in the subtropics in summer, with enhanced stationary eddy momentum flux divergence (SEMFD) when the precipitation and Tibetan High are enhanced in July. This SEMFD in turn drives the negative acceleration of the vertically averaged zonal mean zonal wind $\langle [u] \rangle$ in the subtropics in summer. Therefore, the western edge of the Asian monsoon, and in particular, the intraseasonal change in its strength from June to July appears to be a key process governing the negative acceleration of zonal mean momentum from June to July.
These midsummer variations of $\langle [u] \rangle$ are significantly correlated with both the westward displacement of the NASH and the associated rainfall deficit in the WATL, pointing to the strength of the zonal mean easterlies as an indicator of the strength of the July drying in the WATL. The key steps connecting the Indian monsoon with WATL precipitation are shown in Figure 3.11 along with the respective correlation for each step. Interestingly, the remote correlation between the Indian monsoon and the July drying in the WATL is significant and negative, in addition to the intermediary processes linking the two. The increase in precipitation from June to July in India corresponds to a decrease in precipitation from June to July in the WATL. The role of the monsoon in driving zonal mean easterlies implies that this inverse relationship is not a spurious association, but rather, the Asian monsoon appears to be an important factor governing the variability of midsummer precipitation in the WATL. The findings of this observational analysis lend support to the new hypothesis that the midsummer enhancement of the planetary mean easterlies plays a critical role in the midsummer drying of the subtropical Americas. The westward advection of the NASH’s anticyclonic vorticity across the Atlantic by subtropical easterlies seems key to this drying. Global model experimentation below (Chapter 4) is used to test this mechanistic hypothesis.
Figure 3.11: Summary schematic of the proposed teleconnections linking the Indian monsoon with July drying in the WATL. Implied causation flows from right to left of the page. Linear correlation coefficients of the various processes are shown in the bottom figure with blue arrows indicating direct pathways and red and black arrows indicating indirect connections. Correlations are based on July minus June differences the 1979-2006 period. For wind and SLP data, correlations presented are based on the average MERRA and NCEP-GR1 coefficients.
CHAPTER 4: Asian Monsoon Forcing Experiments

4.1 Model and Methods

Global circulation model (GCM) experiments using version 4 of the Community Atmosphere Model (CAM4) are used to probe the observed linkages presented above and summarized in Figure 3.11. CAM4 is the atmosphere component of the Community Climate System Model version 4 (CCSM4), a fully coupled state-of-the-art climate model for simulating the earth’s climate system (Gent et al., 2011). CAM4 is configured using a finite-volume dynamical core with a 1.9° x 2.5° horizontal resolution and 26 vertical levels. Observed climatological SSTs (Hurrell et al., 2008) act as boundary forcing over ocean grid points, while over land CAM4 is coupled to the Community Land Model version 4 (CLM4; Oleson et al., 2010).

The effects of forcing to the Asian monsoon are explored through a suite of 4 simulations in which the land surface heating over south Asia is steadily increased by changing the soil color. Together with soil wetness, soil color acts to determine the snow-free ground albedo in CLM4. Soil color ranges from 1 to 20, with higher numbers indicating lower reflectance (see Oleson et al., 2010 for more details). Note that the total surface albedo is only partially determined by the soil albedo and in regions of dense vegetation, soil color is essentially ignored in the model. The soil color value is adjusted over the region 50-100°E and 5-40°N (Fig. 4.1). Control values of soil colors in the model (not shown) are quite heterogeneous, ranging from 1-2 (high albedo) over the Middle East to 18-19 (low albedo) in the eastern edge of the domain over eastern Indian/western China. Rather than using this observed color as control, we examine high
soil color (low albedo) minus low soil color (high albedo) for clarity. A total of 4 simulations were carried out in which the soil color is set consistently throughout the boxed domain to a value of 1,3,5, and 20 (SOIL_01, SOIL_03, SOIL_05, and SOIL_20). Each simulation was integrated for 22-years, and the first 2 years were discarded to account for any transient spin-up effects.

Figure 4.1  JJA surface albedo response (%) for (a) WEAK, (b) MEDIUM, and (c) STRONG forcing (grey contours). The region of imposed forcing is indicated by the red dashed line.
Differences between SOIL_20 minus SOIL_05, SOIL_20 minus SOIL_03, and SOIL_20 minus SOIL_01, are respectively called WEAK, MEDIUM, and STRONG experiments. Note that these experiments all represent positive enhancements to the monsoon of varying magnitude. The biggest change in surface albedo is preferentially on the northwest corner of the domain (Fig. 4.1) where vegetation is relatively scarce and hence, soil color plays a more significant role. The domain averaged (land points only) surface albedo, net surface radiation, and surface temperature for WEAK, MEDIUM, and STRONG are shown in Table 4.1. Forcings of +5, +10, and +20 Wm$^{-2}$ in net radiation are the net result of these boundary condition experiments.

Note that modeling results in this chapter examine mean seasonal (JJA) differences of fields due to different monsoon forcings, as opposed to sub-seasonal differences (July minus June) of interannual variability in observations (Chapter 3). From a technical standpoint, this is due to the fact that the aforementioned soil albedo intrusions represent fixed (in time) changes to a boundary condition, so that the monsoon is systematically strengthened throughout the summer season. From a physical standpoint, this is justified since the mechanism linking the monsoon to $\langle [u] \rangle$ and climate impacts in the WATL should in principle act on all timescales. July minus June differences are better suited to detect unique interannual connections in observations driven by $\langle [u] \rangle$ anomalies (see Chapter 3), as these high frequency anomalies have less influence from ENSO and other modes of low frequency climate variability. The modeling experiments conducted here, however, use a climatological SST distribution throughout the integration, so low frequency variations of SST forcing are not a factor in interpreting the JJA response due to atmospheric forcing.
<table>
<thead>
<tr>
<th>Experiment</th>
<th>Δ sfc Albedo (%)</th>
<th>Δ sfc Net Rad (Wm(^{-2}))</th>
<th>Δ sfc Temp (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WEAK</td>
<td>-3.7</td>
<td>5.2</td>
<td>0.2</td>
</tr>
<tr>
<td>MEDIUM</td>
<td>-7.0</td>
<td>10.6</td>
<td>0.8</td>
</tr>
<tr>
<td>STRONG</td>
<td>-14.8</td>
<td>20.9</td>
<td>1.8</td>
</tr>
</tbody>
</table>

Table 4.1: JJA response of surface albedo, net surface radiation, and surface temperature for the indicated experiment averaged over 50-100°E and 5-40° (red box in Fig. 4.1) for land points only.

4.2 Rossby Waves, Eddy Momentum Fluxes, and Mean Easterlies

Following Sardeshmukh and Hoskins (1988), the vorticity equation for a single level in the upper troposphere (200 hPa here) can be approximated as:

\[
\frac{\partial \zeta}{\partial t} + \vec{v}_\psi \cdot \nabla (f + \zeta) = RWS
\]  

(4.1)

where the Rossby wave source (RWS) is defined as:

\[
RWS = - \vec{v}_x \cdot \nabla (f + \zeta) - (f + \zeta) \nabla \cdot \vec{v}_x
\]  

(4.2)
The notation here is standard and the total wind vector $\mathbf{v}$ has been decomposed into its rotational ($\mathbf{v}_r$) and divergent ($\mathbf{v}_\chi$) components. The first term on the right of Equation 4.2 is the advection of absolute vorticity by the divergent wind and the second term is the stretching or divergence term. Hence, the evolution of the vorticity field which describes Rossby waves is driven by the action of the divergent wind which generates them. RWS at 200 hPa is shown in Figure 4.2 along with the 500 hPa omega field for WEAK, MEDIUM, and STRONG cases.

As heating over south Asia is ramped up, there is increasing upward motion near the Indian subcontinent, particularly over the Arabian Sea and the Tibetan Plateau (Fig. 4.2 b,d,f). As expected, there is a broad area of negative RWS and upper-level vorticity divergence (Fig. 4.2 a,c,e) associated with this mid-level ascent. Interestingly, there is a robust area of positive RWS to the northwest of these convective centers, particularly over the Mediterranean near 45°N. The mid-level subsidence over the Mediterranean (Fig. 4.2 b,d,f) corresponding to this upper-level convergence is the “monsoon-desert mechanism” Rodwell and Hoskins (1996) showed on the 325K isentrope. They argue this isentropic downglide occurs as the equatorward flank of the mid-latitude westerlies interacts with the warm thermal structure of the westward propagating stationary Rossby wave response to heating. The positive RWS over the Mediterranean corresponds to the northwest corner of the upper level TH anticyclone (see Figs. 2.4, 2.5). Extratropical convergence forms here as the TH extends westward and interacts with the background westerlies as heating is increased. This finding is consistent with previous studies that examined the RWS response to Asian monsoon-like heating in a 3-D baroclinic atmosphere (Lin, 2009; Jin and Hoskins, 1995; Qin and Robinson, 1993). Although total
RWS is shown here, the positive RWS response in Figure 4.2 is mainly due to the
divergence term in Equation 4.2. This contrasts to the original emphasis by Sardeshmukh
and Hoskins (1988) on vorticity advection by the divergent wind. The main distinction is
that their barotropic model imposed a divergence perturbation a priori, not allowing for
extratropical convergence to develop outside of the specified area of tropical divergence
(heating).

The wave activity flux formulated by Plumb (1985) is an extension of the
the classic Eliassen-Palm relation (Edmon et al., 1980) to three-dimensions, giving
insight into the propagation of longitudinally varying stationary waves. We use wave
activity flux following Plumb (1985) to highlight the propagation of upper-level Rossby
waves away from their source on the northwest corner of the TH. The flux $\mathbf{F}_s$ is parallel
to the local group velocity of planetary waves and is independent of wave phase, as in the
zonally averaged form given by Eliassen-Palm. Figure 4.3 shows the mean summer eddy
vorticity (color) and streamfunction (grey contours) at 200 hPa for each experiment
overlain with $\mathbf{F}_s$ (arrows). Increasing land surface heating (soil_01 $\rightarrow$ soil_20) forces a
stronger upper-level anticyclone centered near 40°E and 30°N as seen in the eddy stream
function field, with a corresponding increase in the vorticity dipole along 40°N. There is a
clear flux of wave activity diverging out of the vorticity maximum (around 20°E, 45°N)
to the southeast over Middle East. This vorticity maximum corresponds to the maximum
in RWS identified in Figure 4.2. There is also a secondary area of wave activity diverging
out of the northeastern flank of the eddy anticyclone north of India, converging over
southeast Asia. This wave activity flux is weaker than the western center because the $\gamma$-
component of $\mathbf{F}_s$ depends heavily on $v^*$. The eastern part of the TH (over Tibet) is
zonally elongated, whereas its western part near the eddy stream function maximum has a greater phase tilt in the meridional plane as the anomalous anticyclone interacts with the oncoming westerlies (Figs. 2.4, 2.5). The meridional propagation of Rossby waves away from their source on the northwest corner of the TH implies momentum convergence there (westerly acceleration), since Rossby waves transport zonal momentum in the opposite direction of their wave activity flux. By the same token, momentum divergence (easterly acceleration) is generated over the Middle East near 30°N as Rossby waves are absorbed near the middle of the TH anticyclone (Fig. 4.3).

Figure 4.2: JJA response of the Rossby wave source 200 hPa (a,c,e) along with the corresponding change in omega at 500mb (b,d,f) due to different forcings.
Figure 4.3: Mean values of the zonally asymmetric component of vorticity (colors) and stream function (grey contours, contour interval = 1e6 * m² s⁻¹) at 200 hPa for each simulation. The wave activity flux following Plumb (1985) is given by arrows.
This connection between Rossby wave propagation and momentum transport is directly confirmed in Fig. 4.4, which shows the impact of increased monsoon heating on stationary eddy momentum fluxes ($u^*v^*$; left panels) and their convergence (right panels). The biggest eddy momentum flux changes (Fig. 4.4 a,c,e) are again on the northwest corner of the TH where the contours of the anomalous anticyclone are tilted almost at 45° (Fig. 4.3), maximizing the product $u^*v^*$. As monsoon heating is increased, the magnitude of momentum convergence (westerly acceleration; red in right panels) and divergence (easterly acceleration; blue in right panels) is increased on the poleward and equatorward flanks of this $u^*v^*$ maximum, respectively (Fig. 4.4 b,d,f). Momentum convergence around 45°N corresponds to the Rossby source region identified in Figure 4.2, and the area of momentum divergence near 30°N over the Middle East is consistent with the propagation of Rossby waves away from this source (Fig. 4.3).

The global response of stationary eddy momentum fluxes is not just confined to the longitudes of the monsoon. Considerable changes in the form of a nearly fixed pattern with amplitude proportional to the forcing are also seen over the eastern oceanic TUTTS for instance. But as summarized in Figure 4.5, local Rossby wave forcing of the mean flow on the western margin of the TH (~25°E) is the most prominent feature driving the zonal mean. The background CAM4 climatology of SEMFC is shown in Fig. 4.5 (colors) in addition to its response to the monsoon heating experiments (lines). Interestingly, the model’s response is largely an amplification of its basic state: the perturbed SEMFC distribution closely follows its climatological distribution, both for the Asian TH as well as the Atlantic and Pacific TUTTS. This reinforces the conclusion of previous studies using simpler models that local heating from the Asian monsoon is the primary control of
the global stationary wave pattern in summer (e.g. Ting, 1994; Hoskins and Rodwell, 1995; Chen, 2001). Also note that CAM4’s mean climatology of SEMFC (colors in Fig. 4.5) is quite realistic (similar to MERRA observations, compare Fig. 4.5 and bottom panel of Fig. 2.5) as the western TH is crucial to zonal mean SEMFC in both. One distinction, however, is that CAM4’s TH circulation (and associated $u^*v^*$) is shifted too far west by about 10-20°, consistent with its westward shifted monsoon precipitation bias (see Figure B.1 and also Meehl et al., 2012).

Figure 4.4: JJA response of 200hPa stationary eddy momentum flux (a,c,e) and eddy momentum flux divergence (b,d,f) due to different forcings. Blue colors in (b,d,f) imply an easterly tendency.
Since ultimately it is changes in the *zonal mean* of Fig. 4.5 that matters to $\langle [u] \rangle$ (see Equation 2.4), Figure 4.6 (a,c,e) shows the zonally averaged response of SEMFC to steadily increasing monsoon forcing. The main feature is the area of negative SEMFC (i.e. divergence) around 30°N in the upper troposphere which drives an easterly acceleration of increasing magnitude as monsoon heating is increased. Also evident is the low-level eddy dipole between 10-20°N, corresponding to the intensification of the famous Indian monsoon southwesterlies or Somali jet (Findlater, 1969a,b), which brings warm, humid air onshore to south Asia. However, Rodwell and Hoskins (1995) show
potential vorticity constraints dictate that this meridional flow drags along East Africa highlands with lateral friction a vorticity sink. Thus eddy \( u^*v^* \) may be partly balanced by frictional and perhaps mountain torques, making low-levels less influential in driving the column mean SEMFC in the subtropics.

Figure 4.6: JJA response of zonal mean eddy momentum flux convergence (a,c,e) and mean meridional stream function (b,d,f) due to different forcings. Blue in (a,c,e) imply an easterly tendency and solid contours in (b,d,f) imply clockwise rotation.
Eddy-induced acceleration of the mean zonal wind at any one level will drive a mean meridional circulation (MMC) to rebalance the mass field to ensure geostrophic (and thermal wind) balance is maintained. Hence, we show changes in the mean meridional stream function field alongside SEMFC in Figure 4.6 (b,d,f). An enhancement of the cross-equatorial thermally direct Hadley cell in the tropics (see for instance Dima et al., 2005) is clearly seen in Figure 4.6 (b,d,f), with circulation changes centered near 10°N and maximum ascent at 25°N. The Coriolis force acting on this Hadley response in the deep tropics (south of 20°N) drives equatorward flow aloft (easterlies) and poleward flow near the surface (westerlies). In the subtropics (centered near 30°N), the summer hemisphere Hadley cell also exhibits changes proportional to the forcing, but in contrast to the cross-equatorial cell, eddy momentum fluxes play a more important role than diabatic heating in driving the summer hemisphere MMC (Schneider and Bordoni, 2008). The easterly torque at upper levels near 30°N (negative SEMFC) in Figure 4.6 (b,d,f) drives mean meridional motions (clockwise circulation) with poleward flowing air aloft offset by equatorward flowing air near the surface. The Coriolis deflection of the equatorward flowing air at subtropical latitudes (Fig. 4.6 b,d,f) implies an easterly response of the zonal wind at low levels.

Changes in \([u]\) and \([\bar{u}]\) can be seen in Figure 4.7. There is a robust equivalent barotropic response of the mean zonal wind in the subtropics centered around 25-30°N, with easterlies increasing in a monotonic (with height) and linear fashion with increased monsoon heating. Peak changes are in the upper troposphere where the maximum in SEMFC forcing acts (Fig 4.6), but with a deep easterly response extending all the way
down to the surface in the subtropics. The compensating mid-latitude westerly response of the mean flow (near 45°N) is also evident, which corresponds to the Rossby wave source region on the northwest corner of the TH (Fig. 4.2).

**Figure 4.7:** JJA response of the mean zonal wind (left panels) along with the corresponding vertically averaged values (right panels) due to different forcings.
4.3 Drying in the Western Atlantic Subtropics

The downstream impacts of a stronger Asian monsoon and zonal mean easterlies on west Atlantic climate are examined here. As seen above, increasing the Asian land-sea temperature contrast creates a more vigorous south Asian monsoon with increased onshore moisture flow and precipitation. The near-surface monsoon cyclone and upper level anticyclone are enhanced (Fig. 4.8) over the Middle East, to the northwest of precipitation (Fig. 4.9) as suggested by the Gill (1980) mechanism. Near-surface anticyclonic anomalies over the Mediterranean and northern Sahara are also consistent with the monsoon-desert mechanism (Rodwell and Hoskins, 1996) discussed previously. Maximum monsoon rainfall changes (Fig. 4.9) are collocated in regions where the mean rainfall is (excessively) large in CAM4’s control climatology, namely over the Arabian Sea and the steep foothills of the Tibetan Plateau (Fig. 4.9d).

The western Atlantic and neighboring land masses systematically become drier (Fig.4.10) as the Asian monsoon gets wetter (Fig. 4.9). Consistent with this drying is the development of near surface anticyclonic anomalies around the Gulf of Mexico (Fig. 4.8). Meanwhile, near surface cyclonic anomalies envelop the subtropical East Atlantic (east of ~ 60°E), an east-west vorticity dipole pattern which resembles the midsummer change in the NASH seen in observed climatology (Fig. 1.4). In accord with changes in $\langle u \rangle$ (Fig. 4.7), the low-level wind over the subtropical Atlantic (~ 20-30°N) becomes increasingly easterly (Fig. 4.10) as the monsoon is enhanced, displacing the southwest corner of the NASH westward (Fig. 4.8). The western enhancement of the NASH and

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4 Precipitation changes over South Asia and the Atlantic are plotted separately in Figures 4.9 and 4.10 due to their different scales.
drying in the western Atlantic subtropics with increased monsoon forcings supports the causal mechanisms suggested in the correlation analysis of Chapter 3. The western edge of the Pacific anticyclone also dries the subtropics where easterly anomalies have been induced, but less.

Another robust change is the increase in rainfall in the East Atlantic ITCZ around 10°N as the monsoon forcing is steadily increased (Fig. 4.10). As the trade winds decrease in the tropical East Atlantic, anomalous westerlies transport moisture from west to east at 10°N. This east-west rain dipole in the tropical Atlantic is also in the same sense of its observed climatology, which shows a rainfall decrease (increase) in the western (eastern) basin from early to midsummer (Fig. 1.2a), so onset and magnitude of the Asian monsoon may also be an important factor in explaining the rainfall climatology of the Atlantic ITCZ.

![Figure 4.8: JJA response of stream function at 1000 hPa (left panels) and 200 hPa (right panels) due to different forcings.](image)
Figure 4.9: (a-c) JJA response of precipitation (colors) and surface wind (arrows) over South Asia due to different forcings. (d) The corresponding mean climatology in the CAM4 control simulation.
Figure 4.10: As in Figure 4.9, except for the Atlantic.
A simple moisture budget is analyzed here to diagnose the source of drying in the western Atlantic. The relevant balance equation assuming a steady state can be estimated as:

\[
\text{precip} - \text{evap} = - \frac{1}{\rho_w g} \int_{SFC}^{TOA} (\vec{v} \cdot \nabla q + q \nabla \cdot \vec{v}) \, dp \tag{4.3}
\]

where \(\vec{v}\) is the horizontal wind vector, \(\rho_w\) is liquid water density, \(g\) is gravity, and \(q\) is specific humidity. The first term in the parentheses is the vertically integrated advection of humidity by the horizontal wind. The second term is the vertically integrated product of moisture and convergence. Together, these two terms constitute moisture flux convergence (MFC). Changes in the total MFC as well as the contribution from low-level wind divergence \((q \nabla \cdot \vec{v})\) are shown in Figure 4.11, where negative values indicate a drying tendency. The large-scale drying in the western Atlantic is predominately due to the low-level wind divergence term \((q \nabla \cdot \vec{v})\) in the westward displaced NASH (Fig. 4.8), as its pattern closely resembles the total MFC. However the advective component \((\vec{v} \cdot \nabla q)\) of MFC by northeasterlies (Fig. 4.10) is a contributing factor locally, for instance the enhancement of dry anomalies in the Bahamas and off the southeastern US, and over Central Mexico (~100W; 20N) where orographic effects matter. Further decomposition of the divergence term \((q \nabla \cdot \vec{v})\) in to the respective contribution from perturbation to the low-level wind and perturbation to humidity (Fig. 4.12) indicates that the large-scale drying across the WATL basin (Fig. 4.11) is primarily due to changes in divergent low-level wind field in the westward displaced NASH. Its worth noting here that the model erroneously (relative to the observations in Fig. 1.2) produces an increase in precipitation along the Pacific coast of southern Mexico due to convergence of the
near-surface wind. Cooling of SSTs by enhanced easterlies trades winds on the southwest corner of the NASH appear fundamental in creating local divergence here and hence, a drying response (similar to observations in Fig. 1.2). This role of local ocean-atmosphere coupling along the Pacific coast is discussed in Chapter 5.2.

Figure 4.11: JJA response of total moisture flux convergence (a,c,e) and the contribution from the wind divergence term \( q \nabla \cdot \mathbf{v} \) (b,d,f) due to different forcings.
Figure 4.12: Decomposition of the moisture tendency due to low-level wind divergence $q \nabla \cdot \vec{v}$ into the contribution from a perturbation to the surface wind field $(\nabla \cdot \vec{v})'$ (a,c,e) and a perturbation to the humidity field $(q)'$ (b,d,f). Where $(\cdot)'$ indicates the model’s JJA response to the various forcings.

4.4 Summary

In a series of experiments with the Community Atmosphere Model version 4 (CAM4), a soil color (albedo) boundary condition is progressively darkened in south Asia (Fig. 4.1), causing net surface radiation changes (including all feedbacks) of about 5, 10 and 20 Wm$^{-2}$ from top to bottom. As a result, rainfall is enhanced in South Asia
(Fig. 4.9), while the western Atlantic is progressively dried (Fig. 4.10), in a pattern indicative of a westward shift of the NASH circulation system (Fig. 4.8), with its dry fine weather, by the monsoon-driven zonal mean wind differences (Fig. 4.7). These changes in the vertically averaged (or barotropic) mean zonal wind are due to changes in stationary eddy momentum fluxes, which act primarily at upper levels near 30°N (Fig. 4.6). Changes in the zonal mean SEMFC at upper levels are primarily due to eddy divergence on the western edge of the upper-level TH (Fig. 4.4), which in turn are due to the southeasterly flux of upper-level Rossby waves out of their source in eastern Europe (Figs. 4.2, 4.3). In summary, monsoon forcing of the upper-level TH anticyclone is a key mechanism governing the enhancement of mean subtropical easterlies in summer. These AGCM results substantiate the mechanistic hypothesis derived from observations above (see Figure 3.11). Large-scale atmospheric dynamics associated with the onset of the Asian monsoon are fundamental to the climatological midsummer drying in the WATL, whereby a sign change of the mean flow causes a westward shift of the NASH. Though since a large portion of the WATL region resides over warm open waters, we next consider the role of local air-sea interactions on the model response.
CHAPTER 5: Air-Sea Interactions in the Western Atlantic

5.1 The Atlantic Warm Pool

The WATL region studied here includes the large body of very warm water known collectively as the Atlantic Warm Pool (AWP), which encompasses the Gulf of Mexico, Caribbean Sea, and western tropical North Atlantic. SSTs are slowly varying and actually peak in the WATL during the July-August dry period (Wang and Enfield, 2003). Hence, simple thermodynamic constraints of the AWP can not readily explain the high frequency precipitation changes of the climatological MSD. However, the AWP varies considerably on interannual and longer timescales (Wang et al., 2006) and has been shown to have a significant impact on summer precipitation in the WATL. Modeling work by Wang et al. (2007, 2008) demonstrate the role AWP variability in modulating summer precipitation in the WATL via its direct effect on sea-level pressure anomalies: A weaker/small (stronger/larger) AWP leads to an increase (decrease) in intensity of the NASH and reduced (increased) precipitation.

AWP and $\overline{[u]}$ variations can affect the NASH in the WATL through different pathways: The AWP acts as a direct source/sink of local vorticity anomalies through heating-induced changes, whereas changes in $\overline{[u]}$ govern the distribution of vorticity anomalies through advective changes across the Atlantic. Non-linear atmospheric dynamics, characterized by a sign change of $\overline{[u]}$, appear fundamental to explain the sub-seasonal changes of the MSD (as detailed above). However, slowly evolving SST
anomalies have a significant impact on summer precipitation in the WATL (Wang et al., 2007) and thus can potentially mediate the impacts from higher frequency atmosphere forcing.

Figure 5.1: Composite SST difference when the SST is anomalously large and small used as boundary forcing.
The STRONG monsoon forcing experiments (Figure 4.1; Table 4.1) are repeated using small and large AWP as the SST boundary condition to estimate what role natural AWP variations play in amplifying or damping the climate response in the WATL from remote atmosphere forcing. Upstream changes on \( \{[u]\} \) and the Asian monsoon (not shown) remain independent of the AWP size, so that WATL climate differences (Figs 5.2 and 5.3) point to changes in the local response under consistent forcing. The size of the AWP has been shown to vary in association with the Atlantic Multidecadal Oscillation (AMO), with large (small) AWPs more frequent during the warm (cool) phase of the AMO (Wang et al., 2006). The AWP is delineated by waters warmer than 28.5°C and the methodology for defining “large” and “small” AWPs follows Wang et al. (2008):

A warm pool 33% larger (smaller) than the climatological warm pool area is identified as a large (small) warm pool; otherwise, warm pools are classified as normal or neutral. From 1949–2001, there are six large warm pools (1952, 1958, 1969, 1987, 1995, 1998) and seven small warm pools (1971, 1974, 1975, 1976, 1984, 1986, 1992) . . . Based on these years for large and small warm pools, we make SST composites for large and small warm pools.

Monthly SSTs from the large and small AWP composites are applied to the AWP region (5-30°N between 40°W and the coast of the Americas; box in Figure 5.1), and the observed climatological SSTs (Hurrell, 2008) are applied to the rest of the globe. A simple smoothing function is applied at the AWP boundaries to prevent discontinuities. The JJA mean SST forcing differences between the small/large AWP runs and climatology are shown in Figure 5.1 Maximum local anomalies exhibit a departure from climatology of around \(+/-\ 0.6°C\). Figure 5.2 shows the SLP response to the STRONG soil color forcing with bottom-boundary forcing from observed climatological SSTs (as before), small AWP, or large AWP added to the SOIL_20 run. Changes in local SSTs associated with the size of
the AWP clearly have a significant affect on the mean JJA SLP response in the WATL. The westward displacement of the NASH is evident in all simulations, but its enhancement in the WATL is considerably increased (decreased) when the AWP is small (large). The basin-wide SLP response in the subtropical Atlantic is more zonally asymmetric when the AWP is small: local forcing by cooler SST’s act together with westward advection by $<u>$ to preferentially enhance the NASH on its western edge.

Figure 5.2: Monsoon-forced SLP difference in the Atlantic using varying SST anomalies in the AWP.
Precipitation and surface wind changes are shown in Figure 5.3. The biggest changes between small and large AWP simulations are seen in the southern portion of the WATL around 10-20°N, consistent with maximum differences in local SST forcing (see Fig. 5.1). The weakening of the NASH due to a larger AWP (warmer SSTs) significantly reduces the meridional pressure gradient in the Caribbean (10-20°N), thereby weakening the Caribbean low-level jet (CLLJ) and its westward moisture transport (Fig. 5.3). A weakening of the CLLJ with a large AWP has a dramatic impact on the precipitation response, causing convergence on the Caribbean side of Central America and hence, a wet rainfall response instead of dry. These changes are in the same sense of previous work which demonstrates the interconnectivity between the AWP, NASH, and CLLJ (Wang and Lee, 2007).

Anomalous SST forcing associated with the extent of the AWP clearly has a significant affect on the distribution and magnitude of the mean summer SLP and precipitation response to monsoon forcing in the WATL. A large (small) AWP dampens (amplifies) the response from using climatological SSTs, particularly in the Caribbean south of 20°N. However, an anomalous AWP does not obfuscate the sub-seasonal changes characteristic of the MSD. Figure 5.4 shows the mean annual cycle of precipitation and SLP averaged over the WATL in CAM4 control simulations coupled to climatological SSTs, a small AWP, and a large AWP. The AWP acts as an important boundary condition, constraining the magnitude of SLP and precipitation, shifting its annual cycle up or down, but leaving its shape nearly unchanged. The MSD and the associated rise in SLP is well-defined regardless of SST forcing. In fact, the MSD is even more defined (in the sense of higher variance) when summer SSTs are greatest: higher
mean summer precipitation due to a large AWP permit a greater midsummer reduction in precipitation as \(\{u\}\) easterlies intrude. Natural variability in AWP size associated with the AMO are important in explaining mean summer rainfall differences in the WATL (Wang et al., 2006; 2007) but these results suggest that SST anomalies are not fundamental to explaining the climatological time signature of the precipitation in the region, with its distinct midsummer drying in July-August.

Figure 5.3: Monsoon-forced precipitation and surface wind difference in the Atlantic using varying SST anomalies in the AWP.
Figure 5.4: CAM4 monthly climatology of (a) precipitation and (b) SLP in the WATL using varying SST anomalies in the AWP region.
5.2 Slab Ocean Model Experiments

The previous section examined the direct forcing of SST anomalies on mediating the local SLP and precipitation response in the WATL. SST variations may also influence the model’s precipitation response indirectly through a wind-evaporation-sea surface temperature (WES) feedback (Xie and Philander, 1994; Xie, 1996): stronger easterly trade winds associated with a stronger NASH lead to greater evaporative cooling (latent heat flux) and lower SSTs, which suppresses convection and may reinforce the NASH and rainfall deficit further. On the other hand, the SST feedbacks in the WATL may be negative: the decrease in cloudiness and rainfall from a westward displaced NASH might increase the insolation, possibly warming the ocean’s surface and destabilizing the planetary boundary layer, thus decreasing the drying response in the WATL. Previous studies indicate that the feedback is generally positive in warm tropical waters where SST forcing is strong (Kumar and Hoerling 1998; Lau and Nath 2000, 2003) but can be negative (implying the atmosphere forces the ocean) in the extratropics where SSTs are cooler (Lau and Nath, 1994). To gain insight on the role of prescribed SST forcing versus coupled air-sea interactions in the WATL, the STRONG monsoon-forcing experiments (Chapter 4) are repeated using identical atmosphere forcing but with CAM4 coupled to a simple slab-ocean model (SOM) instead of fixed SSTs.

SOM runs have only a single layer thermodynamic ocean and sea ice with specified heat fluxes through the bottom. This simple coupling allows for the role of SST feedbacks due to local (1D) air-sea exchanges in the WATL to be addressed. For the CAM4-SOM simulations, the model is integrated for 30 years and only the last 20 years
are used to allow the model to equilibrate. The JJA SLP and precipitation response for the STRONG forcing (Table 4.1) with CAM4-SOM is shown in Figure 5.5, in addition to the fixed SST response repeated here for reference. The WATL SLP response to monsoon-forcing is significantly greater in the SOM run than SSTs, particularly off the coast of Florida. The greater enhancement of the NASH in the SOM also resembles the response seen when cool SST anomalies associated with a small AWP are added (Fig. 5.2). This similarity in SLP between the SOM response and the forced response using prescribed (negative) SST anomalies suggests a positive feedback between the coupled system where ocean cooling intensifies the NASH.

Figure 5.5: JJA SLP, precipitation and surface wind differences for the STRONG forcing using climatological fixed SSTs versus a SOM.
The most striking difference in the precipitation response between the SOM and fixed SST run are off the Pacific coast of southern Mexico (Fig. 5.5). The SOM simulates a pronounced drying near 90°W, 12°N, whereas the fixed SST run (erroneously) shows a wet response here due to low-level wind convergence (Figs. 4.11, 4.12). This is in spite of the fact that both simulations produce an easterly wind response due to the westward displaced NASH. Note that while the small AWP above (cool SST anomalies) in Figure 5.3b does not capture the pronounced drying in the Pacific as in the coupled simulation (Fig. 5.5), this is likely an artifact of the prescribed SST anomalies not extending west of the Atlantic coast of Mesoamerica by design (see Fig. 5.1). Future simulations should explore to what extent SST anomalies capture the coupled response across the global domain.

Changes in the moisture budget terms in the SOM response to STRONG forcing are shown in Figures 5.6-5.8. The large-scale drying of the WATL in the SOM response is primarily due to the divergence term \( \nabla \cdot \vec{v} \) (Fig. 5.6), but the zonal advection of moisture across the Sierra Madre mountains of Mexico is crucial to the enhanced drying along the Pacific coast (Fig. 5.6c). Further decomposition of the divergence term \( \nabla \cdot \vec{v} \) into its respective contribution from perturbations to the wind and humidity field indicate that the drying tendency is due to increased wind divergence rather than a moisture sink (Fig. 5.7). This decomposition of the moisture budget in the SOM is in a similar sense to the results for the fixed SST case (Fig. 4.11).

Figure 5.8 compares moisture budget changes in the SOM directly to the fixed SST run to better understand the rainfall differences in the coupled versus uncoupled case (see Fig. 5.5). The decrease in moisture flux convergence (drying tendency) along the
Pacific coast of southern Mexico in the SOM versus fixed SST response is primarily due to the divergence term \( q \nabla \cdot \tilde{v} \) with the advection term \( \tilde{v} \cdot q \nabla \) playing a secondary role. The increased low-level wind divergence (drying tendency) along the Pacific coast (~ 90°W; 12°N) appears directly related to local cooling of SSTs in the SOM (Fig. 5.9). Figure 5.10 shows the response of the surface energy budget terms in the SOM where fluxes are defined positive into the ocean’s surface. The biggest changes in the WATL are in net shortwave radiation and latent heat flux, both quantities having nearly equal magnitude but of opposite sign. The western enhancement of the NASH causes warming of the ocean’s surface from enhanced insolation (reduced cloudiness) but also cooling as enhanced easterly trades (Fig. 5.5b) create greater evaporation and latent heat flux (Fig. 5.10d). This cooling from evaporation is largest in the eastern Pacific (~ 90°W; 12°N) which corresponds to the greatest precipitation decrease (Fig. 5.5b) and also the greatest cooling of SSTs (Fig. 5.9). Local air-sea feedbacks are largely positive in the WATL, as ocean cooling amplifies the drying driven by large-scale changes in the NASH and near-surface flow. This positive feedback appears particularly important in producing a realistic drying signal off the Pacific coast of southern Mexico in CAM4.

The MSD is strongest off the Pacific coast in observations (Fig. 1.2), similar to the maximum drying signal in the SOM (Fig. 5.5b). This suggests the importance of air-sea feedbacks in capturing the observed drying in the far eastern Pacific ITCZ region. But in contrast to the emphasis by Magaña et al. (1999), the importance of coupled air-sea feedbacks here is on determining the local magnitude of the rainfall response, not in generating the forcing for the MSD itself. Magaña et al (1999) hypothesizes that the MSD is generated locally as cloud cover in early summer reduces insolation and cools
SSTs causing a reduction in precipitation in midsummer. The increase in drying off the Pacific in the SOM, however, is due to enhanced divergence due to evaporative cooling of SSTs, which in turn is due stronger easterly trade winds across Central America as the NASH is displaced west due to easterly $\langle [u] \rangle$ anomalies.

Figure 5.6: JJA response of the moisture budget equation due to STRONG forcing in the SOM run. (a) total moisture flux convergence (b) contribution from $q \nabla \cdot \vec{v}$ and (c) contribution from $\vec{v} \cdot q \nabla$. 


Figure 5.7: Decomposition of the moisture tendency due to low-level wind divergence $q \nabla \cdot \vec{v}$ into the contribution from a perturbation to the surface wind field $(\nabla \cdot \vec{v})'$ (a,c,e) and a perturbation to the humidity field $(q)'$ (b,d,f). Where $(\cdot)'$ indicates the model’s JJA response to STRONG forcing in the SOM run.
Figure 5.8: SOM minus SST model differences of the moisture budget for the JJA response to STRONG monsoon forcing.
Fig 5.9: JJA SST response due to STRONG forcing in the CAM4-SOM simulations.

Figure 5.10: The JJA surface energy budget response over the ocean for STRONG forcing in the SOM. Fluxes are defined positive downward into the ocean’s surface.
CHAPTER 6: Conclusions

The drying in the western Atlantic basin around July and August known as the midsummer drought (MSD) has been investigated in the context of planetary-scale atmospheric dynamics. Robust yet subtle linkages connect the Asian monsoon, planetary zonal mean flow, and western Atlantic summer climate. The MSD in the western Atlantic is fundamentally driven by a westward shift of the circulation regime in the Northern Hemisphere subtropics. This shift is due the protrusion of zonal mean easterly flow into the subtropics in midsummer. Figure 6.1 shows an observed daily climatology of meridional wind $v$ at 850 hPa. The upper map shows the midsummer mean field (colors) with wind arrows, while the central panel is a longitude-time section at 30°N. At right, the vertical mean zonal mean zonal wind climatology $\langle [u] \rangle$ is shown vs. latitude. Notice on Figure 6.1 how the peak of boreal summer, centered on August 1, is characterized by barotropic easterlies extending northward up to 30°N (indicated with heavy pink lines). In the main panel, these easterlies can be seen to cause both westward motion of short waves (seen as fine streaks in this daily climatology), and a westward displacement of the basin-scale circulations over the oceans where they are not topographically anchored. While the US Great Plains low-level jet is anchored (near 100°W), the poleward flow on the western edge of the North Atlantic Subtropical High (NASH) is pressed westward in midsummer. A similar westward press of the Pacific subtropical anticyclone can also be seen (curved dashed annotations).
Figure 6.1: Observed climatology of the meridional wind at 850 hPa ($v_{850}$) and the mean barotropic zonal wind $\langle [u] \rangle$. Top panel shows a map of $v_{850}$ in northern summer overlain with wind vectors. Bottom left panel shows a time-longitude plot of $v_{850}$ at 30ºN with (subjectively drawn) dashed lines indicating the westward displacement of the oceanic subtropical anticyclones. $\langle [u] \rangle$ climatology is shown in the lower-right panel.
Figure 6.2: Schematic of the causal chain of physical processes (right to left) linking the Asian monsoon to summer climate in the western Atlantic and North America. $\langle u \rangle$ is the barotropic zonal mean zonal wind, driven by stationary eddy momentum flux divergence (SEMFD), which displaces the North Atlantic Subtropical High (NASH) westward in midsummer.

The physical connection between the westward shift of the NASH due to easterly $\langle u \rangle$ in midsummer is confirmed in global atmosphere model experiments. These model experiments also examined both the upstream cause of easterly $\langle u \rangle$, as well as the downstream impacts of the westward displaced NASH on summer climate in the western Atlantic. The following conclusions emerge and are also schematically shown in Figure 6.2:

1. Asian monsoon precipitation (heating) drives the upper-level Tibetan High (Fig. 4.8)
2. Stationary eddy meridional momentum flux (SEMFD) on the western edge of the Tibetan High is the primary cause of easterly zonal mean barotropic wind $\langle u \rangle$ in midsummer in the Northern Hemisphere subtropics (Figs. 2.5, 4.4).
3. The easterly $[\mu]$ in the subtropics causes a westward displacement of the North Atlantic Subtropical High (NASH) in midsummer (Fig. 6.1, 4.7, 4.8)

4. The westward displaced NASH induces enhanced low-level wind divergence in the western Atlantic basin (Fig. 4.12)

5. The low-level wind divergence in the western Atlantic causes moisture flux divergence (Fig. 4.11)

6. Moisture flux divergence in the western Atlantic causes reduced precipitation (Fig. 4.9)

7. Local evaporative cooling of sea surface temperatures along the Pacific coast of Central America enhances drying due to low-level wind divergence (Figs. 5.8-5.10)
CHAPTER 7: Future Work: Toward a Unified Understanding of the MSD

While the MSD in the western North Atlantic basin is emphasized above, other regions around the globe experience a MSD too. The southeast corner of Brazil for instance is situated geographically on the western edge of the South Atlantic subtropical anticyclone and experiences a similar midsummer drying in February (Fig. 7.1). The high frequency (HF) precipitation and SLP anomaly (annual and semiannual harmonics removed) centered on February calculated from CMAP climatology is shown on the left panels in Figure 7.1. On the right, the annual cycle of total precipitation and SLP for the box over southeast Brazil is shown, with the time axis shifted 182 days for easier comparison to the Northern Hemisphere. The sharp precipitation decrease in February (austral midsummer) in the western basin of the South Atlantic is analogous to the North Atlantic MSD in July and August (see Fig. 1.2), though the MSD near Brazil is of limited spatial extent and the corresponding SLP rise is more modest.

There is evidence of an east-west dipole of SLP anomalies on the equatorward flank of the anticyclone near 20-30ºS (Fig. 7.1c), suggesting the same \(\langle [u] \rangle\) mechanisms as above may be operating here. Since the connection between the MSD and \(\langle [u] \rangle\) detailed above is inherently a global mechanism, it could in principle act elsewhere. Indeed, the \(\langle [u] \rangle\) climatology in the Southern Hemisphere (Fig 7.2) shows a similar midsummer protrusion of mean tropical easterlies into higher latitudes. Across 15-30ºS (the latitude of the SE Brazil box in Fig. 7.1), there is a sign change of \(\langle [u] \rangle\) from January to February, similar to the June to July change in the Northern Hemisphere subtropics.
(Fig. 1.5). Interannual evidence also indicates that changes in the mean barotropic wind correlates to drying off of Brazil in February. Figure 7.3 shows a scatter plot of February minus January differences between 1979-2005 of CMAP precipitation in the SE Brazil box and $\langle [u] \rangle$ at 15-30°S for both MERRA and NCEP-GR1. There is a consistently positive correlation of about 0.6 across both panels, a particularly strong relationship given how distantly connected these field are.

Figure 7.1: High frequency variations (annual and semiannual Fourier harmonics removed) centered on February for (a) CMAP precipitation (b) and NCEP reanalysis SLP. Right panels show the total time series of each field averaged over the SE Brazil box shown at left. Black contours in (c) shown the climatological mean SLP field in DJF.
Might the easterly \(\langle [u] \rangle\) change in summer govern the MSD in the western South Atlantic as it does in the North Atlantic? As a preliminary test, a modeling experiment was performed in which an additional tendency term is added to the zonal wind equation in CAM4. This \([u]\) forcing is centered at 200 hPa and 20\(^\circ\)S and is applied to all longitudes. Its spatial distribution is Gaussian in latitude and pressure with a maximum magnitude of -7 m s\(^{-1}\) mo\(^{-1}\) when zonally and vertically averaged. The forcing is centered on austral summer (DJF) and smoothly decays away in time. This experiment with this zonal wind perturbation is called UTORQUE-SH and Figure 7.4 examine mean DJF differences of UTORQUE-SH against CAM4’s control climatology (CTL).

The DJF SLP and precipitation response is shown in Figure 7.4 alongside the \(\langle [u] \rangle\) change. Although the forcing is only applied at upper levels, there is a significant easterly response of the barotropic \(\langle [u] \rangle\) flow of about -3 m s\(^{-1}\) at 20\(^\circ\)S due to mass adjustment by the MMC (as discussed in Chapter 2.2). The model’s mean summer precipitation and SLP response is quite similar to the HF anomalies in February in observations (Fig. 7.1). There is drying on the southeast corner of Brazil around 20\(^\circ\)N where the western edge of the South Atlantic subtropical high (SASH) intrudes. The distribution of the SLP change also exhibits a northwest-southeast phase tilt, indicative of a westward displacement on the northwest corner of the SASH driven by the easterly \(\langle [u] \rangle\) response at 20\(^\circ\)S (Fig. 7.4). Its also worth noting that the zonal low-level over the South Atlantic changes in accord with \(\langle [u] \rangle\), both in sign and magnitude, with easterlies of 2-3 m s\(^{-1}\) at 20\(^\circ\)S and westerlies approaching 1 m s\(^{-1}\) near 40\(^\circ\)S.
Figure 7.2: $\langle [u] \rangle$ climatology in the Southern Hemisphere calculated from MERRA from 1979-2005. Time axis is shifted by 182 days for easier visual comparison with the Northern Hemisphere.
I hypothesize that the MSD in the South and North Atlantic can be interpreted in an analogous dynamical framework. Planetary dynamics driven by a sign change of $\langle [u] \rangle$ appear fundamental to the high frequency climate variations in both basins. To make this analogy clearer, an identical experiment to UTORQUE-SH was performed, except with the $[u]$ forcing centered at latitude 25°N in JJA (‘UTORQUE-NH’). The North Atlantic SLP and precipitation response is shown in Figure 7.5 alongside the $\langle [u] \rangle$ change for UTORQUE-NH. The response in Figure 7.5 is close to a mirror image to the Southern Hemisphere case (Fig. 7.4). Both the profile and magnitude of the $\langle [u] \rangle$ change in the northern subtropics closely resembles its Southern Hemisphere counterpart with mean easterlies of about -3 m s$^{-1}$ near 20° latitude (Note the forcing is centered at latitude 25°N in TORQUE-NH in contrast to 20°S in UTORQUE-SH). The NASH is displaced westward on its southwest corner equatorward of 40°N where mean easterlies protrude. The resulting dipole SLP pattern on the equatorward flank of the NASH (Fig. 7.5b).
resembles its observed midsummer climatological anomaly (Fig. 1.4b) and is also analogous to the response of the SASH in UTORQUE-SH (Fig. 7.4b).

Another striking similarity between Figures 7.4 and 7.5 is the increase in eastern basin precipitation on the equatorward flank of the subtropical anticyclones near 10° latitude. Near-surface westerlies associated with cyclonic anomalies generate moisture flux convergence off the west coast of Africa, enhancing the eastern Atlantic ITCZ in both hemispheres and seasons. In the northern WATL, there is drying where the NASH intrudes, particularly off the coast of Florida and the western edge of Mexico. On the eastern edge of Mexico, there is an increase in upslope precipitation where easterlies advect moist air up the Sierra Madre Oriental mountain range.

The SLP and precipitation response in the WATL in UTORQUE-NH (Fig. 7.5) is quite similar to that in the monsoon forcing experiments of Chapter 4 (Figs. 4.8, 4.10). It appears to matter little to the climate response in the WATL whether ⟨[u]⟩ is forced directly through an artificial tendency in the momentum budget equation, or indirectly through thermal forcing of the Asian monsoon. This provides further confirmation that the western enhancement of the NASH and the MSD are controlled by simple advective changes in summer, whereby ⟨[u]⟩ acts as a sort of bridge linking rainfall anomalies over Asia with that over North America.

The preliminary modeling results highlighted in Figures 7.4 and 7.5 indicate that the MSD in the South and North Atlantic are both due to subtle changes of the mean flow in summer and can be interpreted in a similar dynamical framework. Future work will probe the role of ⟨[u]⟩ and climate impacts in different regions as well as different models more extensively. For instance, the implications of the monsoon-forced ⟨[u]⟩ experiments
of Chapter 4 on the western Pacific climate should be explored further. There is a western enhancement of the Pacific subtropical high at 30°N in the East China Sea due to easterly $\langle [u] \rangle$ in both observed climatology (Fig. 6.1) and CAM4’s response to monsoon-forcing (Fig. 4.8). The mechanisms discussed in this work most likely have direct implications on the midsummer rainfall pattern in this region as well. Future work should also entail investigating how the monsoon responds to $\langle [u] \rangle$, i.e. how the zonal mean acts as a feedback on monsoon dynamics, not just a consequence. A complete global survey of summer climate in western ocean basins will be conducted to examine to what extent climate impacts there are due to subtle, yet significant, variations of the mean barotropic flow.

Figure 7.4: (a) JJA Precipitation and surface wind response and (b) JJA SLP response in UTORQUE-SH experiment. Right panels show the $\langle [u] \rangle$ response and black contour in (b) indicates the climatological mean SLP field in the control.
Figure 7.5: As in Fig. 7.4, except for the UTORQUE-NH experiment.
APPENDIX A: Consistency with an Upper Level Heat Source

As a cross-check to the boundary forcing experiments in Chapter 4, monsoon heating experiments using an upper level heat (divergence) source are performed to confirm that the Northern Hemisphere summer climate impacts are not particularly sensitive to how CAM4 is modified. Here we add an extra tendency term to the dry static energy equation in the model’s physics package. The horizontal and vertical distribution of this forcing term is given in Figure A.1. It has a Gaussian horizontal distribution centered at 70ºE, 15ºN, which corresponds to the western sector of the Asian monsoon over the Arabian Sea where CAM4’s JJA rainfall maximum is located (see Fig. 4.9d). The vertical distribution resembles a half-sine wave with a maximum magnitude of +/- 4 K d^{-1} at 400 hPa (Fig. A.1b). A set of two simulations are conducted in which a positive (POS) or negative (NEG) tendency is added in summer. Each simulation is coupled to climatological SSTs and integrated for 22-years with the first 2 years discarded. The following figures show the JJA difference between POS minus NEG.

The stream function response at both 1000 hPa and 200 hPa is shown in Fig A.2. Changes over monsoon region are broadly consistent with Fig. 4.8 in the sense that positive heating anomaly over South Asia drives a low-level cyclonic and upper-level anticyclonic flow to the northwest due to a Rossby wave response (Gill, 1980). The enhancement of the near-surface cyclone and upper-level Tibetan High are displaced further to the west in Figure A.2 (compared to Fig. 4.8), however, consistent with the forcing which is also located further west over the Arabian Sea (Fig. A.1 versus Fig. 4.1). Over the Atlantic, there is an east-west dipole response of the 1000 hPa stream function,
indicating a westward displacement of the NASH very similar to that seen in Fig. 4.8.

The upper-level enhancement of the Tibetan High also causes a marked increased in SEMFD on its western edge near 30ºN (Fig. A.3), similar to the results shown Figure 4.4.

Figure A.1: Distribution of heating anomalies used as forcing in the POS and NEG simulations. Forcing is centered on JJA and smoothly decays away in time.
Zonal mean SEMFC, the mean meridional stream function, and the $[u]$ response are shown in Figure A.4. All the principle connections discussed in detail in Chapter 4 hold when the Asian monsoon is forced by an upper-level heat source. SEMFC aloft near 30ºN forces a counterclockwise MMC response which by deflection by the Coriolis force, drives deep tropospheric easterlies around 30ºN. The low-level westerly tendency by the monsoon southwesterlies is also seen. Values in the Figure A.4 are plotted on the same contouring interval as Figures 4.6 and 4.7 for direct comparison. The magnitude of the response closely resembles the MEDIUM results in from the boundary forcing experiments (middle panel in Figures 4.6 and 4.7).

Rainfall differences in the Atlantic and South Asia are shown in Figure A.5 alongside $\langle [u] \rangle$ changes. Ultimately, a net forcing of $\langle [u] \rangle$ of negative 1-2 m s$^{-1}$ between 20-40ºN (Fig. A.5) displaces the NASH westward (Fig. A.2) causing drying in the WATL subtropics (Fig. A.5) due to low-level wind divergence in the high. Forcing the Asian monsoon through an upper-level heating anomaly on its western edge (Arabian Sea) yield similar results to making the South Asian land surface hotter via lower soil albedo. This consistency points to a robust dynamical mechanism in CAM4 and serves to verify the Northern Hemisphere summer climate linkages discussed above and summarized in Figure 6.2.
Figure A.2: JJA stream function response at (a) 1000 hPa and (b) 200 hPa due to an upper level heating anomaly over the Arabian Sea (POS minus NEG)
Figure A.3: JJA response of 200 hPa (a) stationary eddy momentum flux (b) SEMFC due to an upper level heating anomaly over the Arabian Sea (POS minus NEG)
Figure A.4: JJA zonal mean response of (a) SEMFC (b) streamfunction and (c) zonal wind due to an upper level heating anomaly over the Arabian Sea (POS minus NEG)
Figure A.5: JJA response of precipitation and surface wind (left panels) and $\langle |u| \rangle$ (right panel) due to an upper level heating anomaly over the Arabian Sea (POS minus NEG).
APPENDIX B: Understanding Mean Summer Biases in CAM4

The causal chain linking the Asian monsoon to western Atlantic summer climate deduced from observations (Chapters 2 and 3) and confirmed in dedicated modeling experiments (Chapter 4 and Appendix A), also appears active in CAM4’s mean summer bias in the Northern Hemisphere. Key features of the summer climatology of CAM4 control simulations are discussed here in light of the linkages above. Figure B.1 shows the mean summer (JJA) difference of CAM4 control simulations referenced against CMAP precipitation and MERRA SLP data. All data sets are re-gridded to consistent resolution before differencing.

CAM4 has a large westward bias of Asian monsoon precipitation (see also Meehl et al., 2012), with copious rainfall rates in the Arabian Sea and into southern Arabian peninsula (Fig. B.1) The monsoon low or trough is displaced to the northwest near Israel, consistent with this heating bias. Meanwhile, the monsoon over southeast Asia and the western Pacific is too weak, with dry biases associated with high SLP anomalies where the Pacific subtropical high intrudes. SLP biases over the Atlantic are characterized by an east-west dipole in the subtropics, reminiscent of monsoon-forced changes (Fig. 4.8) and of climatological midsummer SLP changes in observations as the mean flow goes easterly (Fig. 1.4).

The westward biases of Asian monsoon precipitation and westward shift of the ocean subtropical highs are indicative of planetary scale circulation errors in CAM4 which are characterized by an easterly bias of the mean barotropic flow \( \langle u \rangle \). Figure B.2 shows the annual cycle of the mean zonal wind \( \langle u \rangle \) profile in CAM4 and MERRA.
CAM4 has a grossly similar $\langle [u] \rangle$ profile as MERRA but with a considerably stronger midsummer enhancement of the subtropical easterlies that extend too far poleward. Both CAM4 and MERRA have a July minimum in the mean zonal wind $\langle [u] \rangle$, but subtropical easterlies are too strong and also occur too early in CAM4. The monthly time series of rainfall and SLP in CAM4 and MERRA is shown in Figure B.3. CAM4 has a high and dry JJA bias in the WATL consistent with its easterly $\langle [u] \rangle$ bias. Moreover, the onset of the MSD occurs too early in CAM4 (June instead of July), also consistent with its $\langle [u] \rangle$ time series, with mean easterlies occurring too early in June (Fig. B.2). This provides evidence that the physical connection between $\langle [u] \rangle$, the NASH, and rainfall in the WATL is operating on both seasonal and sub-seasonal time scales in the model (also see Appendix C).

The JJA easterly $\langle [u] \rangle$ bias is consistent with too much zonal mean SEMFC near 30°N in CAM4 (Fig. B.4), which appears due in part to errors in the TH (not shown) which extends too far west into the Atlantic in CAM4 (consistent with its rainfall errors in Fig. B.1). The findings from this work suggest that fixing precipitation biases in the Asian monsoon might improve biases at all longitudes in the northern subtropics due to the strong influence of the monsoon on $\langle [u] \rangle$. Such a unified view of the Northern Hemisphere climate system might affect priorities in model development or research directions.
Figure B.1: JJA biases of precipitation and SLP in CAM4 simulations
Figure B.2: Mean $\langle [u] \rangle$ climatology in MERRA and CAM4.
Figure B.3: Time series of (a) SLP and (b) precipitation in the WATL region in CAM4 (blue lines) and observations (black lines).
Figure B.4: JJA climatology calculated from 1979-2005 of zonal mean SEMFC in (a) CAM4 (b) MERRA and (c) their difference.
APPENDIX C: CAM4 Internal Variability

Unforced internal atmospheric variability in CAM4 (with climatological SST boundary condition) is shown to exhibit similar interannual connections as deduced from observations (Chapter 3), lending further support that that the physical processes linking the Asian monsoon with \( \langle [u] \rangle \) and rainfall in the WATL basin are dominated by atmospheric dynamics and are unrelated to ENSO. Figure C.1 shows midsummer rainfall differences composited on various relevant indices in CAM4, constructed in a similar fashion as the observational case (Fig. 3.6-3.7). Note that here, the definitions of the corresponding indices in CAM4 vary subtly from observations to account for mean model biases. For instance, we are looking interannual variations of July minus May rainfall (as opposed to July minus June in observations) since the onset of easterly \( \langle [u] \rangle \) and the MSD in the WATL occurs too early in CAM4 (see Figs. B.2 and B.3).

Furthermore, instead of using the Indian monsoon index (Wang et al., 2000) as a measure of monsoon intensity, we use rainfall over the domain 40-80°E, 10-30°N, which accounts for CAM4’s westward shifted rainfall bias over the Arabian Sea (see Fig. B.1).

CAM4’s internal atmospheric variability tells a similar story to observations: Wet anomalies on the western margin of the monsoon correspond to dry anomalies in the western Atlantic basin (Fig. C.1). The magnitude of the July drying in the WATL is also consistent across the various indices (panels in Fig. C.1). Scatter plots of individual model years of these key indices are also shown in Figure C.2. The increase in precipitation in July is correlated with the easterly acceleration of \( \langle [u] \rangle \) at 20-30°N (Fig. C.2a). The increase in easterly \( \langle [u] \rangle \) is correlated with the drying in the WATL in July.
(Fig. C.2b). And finally, increased monsoon precipitation in July is correlated with decreased precipitation in the WATL (Fig. C.2c). All correlation values are significant at the 95% level and note also, that the strength of these relationships are on a similar scale as their observational counterpart (see Fig. 3.11). These results provide further evidence that the Northern Hemisphere summer climate linkages discussed in this work, operate on both seasonal (e.g. JJA) and sub-seasonal (e.g. July minus June) timescales in the model. Moreover, the fact that the interannual variations of observed midsummer rainfall changes (Fig. 3.6) can be reproduced in CAM4 without any interannual SST variability (Fig. C.1), confirms that internal atmospheric variability, not ENSO, is the driving mechanism.
Figure C.1: July minus May precipitation composite differences based on various indices in CAM4 simulations with climatological SST forcing. Only grid points at the 95% confidence level (using a $t$-test) are shown. Data are coarsened to 10° resolution. See text for explanation of indices.
Figure C.2: July minus May differences based on various indices in CAM4 simulations with climatological SST forcing. Each data point indicates an individual model year. Rainfall data is in units mm/day and $\langle |u| \rangle$ is in units m/s. Pearson’s correlation coefficient $r$ is indicated in each panel.
APPENDIX D: Implications for Paleoclimate Signatures

A robust physical connection should apply to all timescales. This work identifies the connections between the mean barotropic zonal wind $\langle [u] \rangle$, the Asian monsoon, and subtropical climate in western ocean basins on both the sub-seasonal (e.g. July minus June change) as well as seasonal mean (JJA) timescales. It may also have implications for interpreting changes on the paleoclimate timescale as well. As a first glance, Figures D.1-D.3 shows JJA climate model differences between 6kbya (mid-Holocene) and present day. Data from the three climate models shown here are from the Paleoclimate Modelling Intercomparison Project phase 2 (PMIP-2; Crucifix et al., 2005). The mid-Holocene was a period of enhanced insolation and stronger continental monsoons with increased onshore flow. This can be see by JJA rainfall differences in Figure D.1, which generally show positive rainfall differences over land masses in the Northern Hemisphere subtropics. The clearest differences are North Africa which shows a particular strong amplification of the monsoon.

The effects of increased continental heating (due to increased insolation) in the mid-Holocene on the lower tropospheric circulation can be inferred from Figure D.2, which shows JJA changes in eddy geopotential height at 850 hPa. There are lower heights over the Eurasian and North American continents, and over the subtropical Atlantic and Pacific ocean basins, there is evidence of a western intensification of the subtropical highs. This western intensification of the subtropical highs is broadly consistent with the westward displacement seen in present day climatology (Fig. 6.1) as well as monsoon-forced changes (Fig. 4.8) as mean $\langle [u] \rangle$ easterlies protrude in summer.
In fact, mid-Holocene minus present day JJA differences of $\langle u \rangle$ (Fig. D.3) show easterly changes of 1-2 m s$^{-1}$ peaking at 30°N. It's worth mentioning that these $\langle u \rangle$ anomalies are on the same scale as the CAM4 monsoon forcing experiments of Chapter 4 (see Fig. 4.7).

In light of the present analysis, future work should systematically investigate the possible connection between monsoons, $\langle u \rangle$, and subtropical circulation changes in paleoclimate simulations in more detail.

Figure D.1: JJA differences of the precipitation for mid-Holocene (6kbya) minus present day simulations in various PMIP-2 climate models.
Figure D.2: As in Fig. D.1, but for the eddy geopotential height field at 850hPa.
Figure D.3: As in Fig. D.1, but the zonal mean barotropic wind ($[u]$).
APPENDIX E: Implications for Tropical Cyclone Tracks

Advective impacts from \( \langle [u] \rangle \) variations may manifest in other areas beyond MSD research. For instance, \( \langle [u] \rangle \) driven shifts of the subtropical anticyclones may also affect the steering of tropical cyclones. As an example, Figures E.1 and E.2 shows a long-term composite difference of tropical storm days in 10x10° grids (from the IBTrACS data set; Knapp et al., 2010) based on upper minus lower tercile values of \( \langle [u] \rangle \) during the August-October (ASO) storm season. Red-blue anomalies are juxtaposed with a map of storm tracks for reference. The ratio of landfalls to “fish storms” that recurve in the Atlantic and Pacific is clearly affected strongly. Little impact on basinwide storm days is seen (red and blue areas are nearly balanced), suggesting this is not just an ENSO signal (which affects \( \langle [u] \rangle \) but also strongly modulates basinwide numbers). This result is for 20-30°N during tropical storm season (ASO); other latitude belts (e.g. 10-40°N) yielded similar results. While 20-30°N \( \langle [u] \rangle \) in ASO may not be governed solely by the Asian monsoon mechanisms above, Figures E.1-E.2 serve to show the power and relevance of just a few m s\(^{-1}\) difference in this deep layer mean flow component.
Figure E.1: North Atlantic tropical cyclone count differences (in the IBTrACS cyclone database), based on a composite of high minus low tercile \([l_u]\) days in 20-30°N during August-October (ASO). (left panel). Image of tropical cyclone tracks for reference (right panel).

Figure E.2: As in Fig. E.1, but for the entire Northern Hemisphere.
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