Changes in the structure and propagation of the MJO with increasing CO$_2$

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Abstract.

Changes in the Madden-Julian Oscillation (MJO) with increasing CO$_2$ concentrations are examined using the Goddard Institute for Space Studies Global Climate Model (GCM). Four simulations performed with fixed CO$_2$ concentrations of 0.5, 1, 2 and 4 times pre-industrial levels using the GCM coupled with a mixed layer ocean model are analyzed in terms of the basic state, rainfall and moisture variability, and the structure and propagation of the MJO.

The GCM simulates basic state changes associated with increasing CO$_2$ that are consistent with results from earlier studies: column water vapor increases at $\sim 7.1\%\ K^{-1}$, precipitation also increases but at a lower rate ($\sim 3\%\ K^{-1}$), and column relative humidity shows little change. Moisture and rainfall variability intensify with warming. Total moisture and rainfall variability increases at a rate that is similar to that of the mean state change. The intensification is faster in the MJO-related anomalies than in the total anomalies, though the ratio of the MJO band variability to its westward counterpart increases at a much slower rate. On the basis of linear regression analysis and space-time spectral analysis, it is found that the MJO exhibits faster eastward propagation, faster westward energy dispersion, a larger zonal scale and deeper vertical structure in warmer climates.
1. Introduction

The Madden-Julian Oscillation [Madden and Julian, 1971, 1972] is one of the most distinct and prominent modes of tropical intraseasonal (30-90 day) variability [Zhang, 2005; Lau and Waliser, 2011]. It is characterized by an envelope of increased convection several thousand kilometers across, co-located with regions of anomalous tropospheric water vapor. The convective anomalies are coupled to planetary-scale circulation anomalies that resemble the Matsuno-Gill [Matsuno, 1966; Gill, 1980] response to an equatorial heat source. This large-scale couplet of circulation and convection propagates eastward at $\sim 5 \text{ m s}^{-1}$, with widespread influences on weather patterns across the globe [Zhang 2013 and references therein]. As a result, it is of global interest to understand the fundamental dynamics that drive the MJO and be able to predict its behavior.

The topic of how increasing tropospheric temperatures in association with increasing carbon dioxide (CO$_2$) concentrations affect the MJO has gathered more attention throughout the years. Slingo et al. [1999] found that MJO activity is highly dependent on tropical sea surface temperatures, where higher MJO activity occurs when SSTs are higher. They hypothesized that the MJO may strengthen as greenhouse gas concentrations increase. Jones and Carvalho [2011] found that MJO activity has been increasing throughout the 20th century, and estimated that MJO activity will increase by $\sim 50\%$ by the middle of the 21st century. Subsequent modeling studies have found more intense MJO activity as the climate system warms [Liu et al., 2013; Arnold et al., 2013, 2015; Subramanian et al., 2014; Chang et al., 2015; Song and Seo, 2016; Carlson and Caballero, 2016].
While all these studies have found a link between MJO intensity and activity and increasing greenhouse gases, the physical mechanisms under which the MJO intensifies remain poorly understood. While many studies have shown that the MJO exhibits increased eastward propagation [Arnold et al., 2013, 2015; Chang et al., 2015], a quantitative explanation based on theory has not been offered. Furthermore, the relationship between the MJO and the climatological mean state has not been carefully analyzed.

In this study, which comprises two companion papers, we will seek to further our understanding of how the structure and propagation of the MJO changes with increasing CO$_2$ concentrations using linear analysis and theoretical consideration. We will make use of simulations with different CO$_2$ concentrations from the NASA Goddard Institute of Space Studies (GISS) global climate model (GCM) as the main tool of analysis. In this first part, we document the changes in the climatological mean state, tropical variability and MJO structure and propagation. A companion paper [Adames et al., 2017] analyzes changes in the MJO maintenance and propagation through a comprehensive analysis of moist static energy, the moisture-precipitation relationship and a quantitative analysis from the perspective of moisture mode theory (e.g., Sobel and Maloney 2012, 2013; Adames and Kim 2016)

This paper is structured as follows. The following section describes the model and the method of analysis. Section 3 covers the changes in the model mean state with increasing CO$_2$. Section 4 describes changes in tropical variability with increasing CO$_2$. Section 5 documents changes in MJO structure and propagation with increasing CO$_2$. A concluding discussion is offered in Section 6.
2. Data and Methods

2.1. GISS model simulation

To investigate the effect of CO$_2$ changes on the characteristics of the MJO, an atmospheric GCM coupled with a mixed layer ocean model is used. The atmospheric GCM is the atmospheric component of the NASA GISS Model E2 [Schmidt and Coauthors, 2014]. Since GISS Model E2 was used for CMIP5 [Taylor et al., 2012], several important modifications have been made in the model parameterizations, which were shown to improve the model performance. Specifically, changes were made to the convective parameterization toward an increased sensitivity of simulated convection to environmental humidity [Kim et al., 2012; Del Genio et al., 2012]. These changes, which include stronger lateral entrainment rate and convective rain re-evaporation, led the GISS Model E2 to be able to simulate an improved MJO relative to the CMIP5 version [Kim et al., 2012; Del Genio et al., 2012]. The so-called “post-CMIP5” version of the GISS GCM was formulated by incorporating these changes to the convective parameterization [Kim et al., 2012] as well as changes to the convective downdraft and stratiform clouds that mitigate the typical degradation of the mean climate that often accompanies parameterization changes that produce an MJO [Del Genio et al., 2012]. Recent improvements to the boundary layer scheme [Yao and Cheng, 2012] were also included. The post-CMIP5 version of NASA GISS GCM is employed in the current study.

For the mixed layer ocean model, the ocean heat convergence was first obtained from a fixed-SST simulation, in which observed SST was prescribed as the boundary condition. The ocean heat convergence is then prescribed to the mixed layer ocean model in a series of long-term simulations, in which CO$_2$ concentrations were varied. Four simulations with
0.5, 1, 2, and 4 times the current level of CO$_2$ were conducted at a 2.5° (longitude) × 2.0° (latitude) × 40 (levels) resolution, and with a depth of the mixed layer 65 m. After reaching an equilibrium (i.e., steady global mean surface temperature), each experiment was carried out for additional 30-50 years, and the daily-mean data from the last 20 years of simulations are used in the current study.

We make use of the following model output fields in this study. The horizontal wind components ($u$, $v$) are used as field variables, as well as vertical velocity ($\omega$). Specific humidity ($q$), temperature ($T$) and geopotential height ($Z$) are also used as field variables and in the calculation of dry and moist static energies. Temperature is also used in the calculation of the saturation specific humidity ($q_s$) and relative humidity ($q/q_s$).

Other variables that we use include precipitation ($P$), and outgoing longwave radiation (OLR). The Eulerian temporal tendency of moisture, $\partial q/\partial t$ is calculated by taking a 2-day centered difference of $q$.

Many of the fields described in this study correspond to intraseasonal anomalies, obtained by removing the mean and first three harmonics of the annual cycle based on the 20 year simulation. Additionally, a 101-point Lanczos filter [Duchon, 1979] is used to retain anomalies within the 20-100 day timescale. Some of the fields are “MJO-filtered” anomalies, which, in addition to being filtered over the 20-100 day timescale, they are filtered to retain only the eastward-propagating, zonal wavenumber 1-5 signal, based on the protocol of Hayashi [1981].

2.2. Methods

Many of the results shown in this study are obtained through linear regression analysis, obtained following the methods described in Adames and Wallace [2014], which makes
use of the following matrix operation

\[ D = S \hat{P}^T / N \]

where \( D \) is the linear regression for a two-dimensional matrix \( S \), corresponding to a field variable \( S \). The matrix \( S \) is two-dimensional since the field variable \( S \) is reshaped so that its spatial dimensions are contained in the rows of the matrix \( S \) [i.e. \( S = S(x \times y \times p, t) \)].

The regression map is obtained by projecting \( S \) upon a standardized MJO index, \( \hat{P} \), and dividing by the number of days, \( N \). The regression patterns shown later (e.g. Fig. 8) correspond to linear regressions upon an MJO-filtered time series of OLR, averaged over the western Pacific basin (15°N/S 140°-180°E). This region corresponds to the location of strongest intraseasonal variability in the model simulations. We have verified that the results presented in this study are reproducible using time series corresponding to other regions of the Indo-Pacific warm pool (60°E-180°) as well as by using EOF analysis.

Contour and shading intervals in the plots presented here are scaled to the approximate value of the 95% confidence interval based on a two-sided t-test.

Space-time spectral analysis is performed in Section 4 of this study, following the methods of Wheeler and Kiladis [1999], Masunaga [2007], and Hendon and Wheeler [2008]. Time series of precipitation and column water vapor are divided into 180-day segments that overlap by 90 days, as in Waliser et al. [2009] and Kim et al. [2009]. Then, the space-time mean and linear trends are removed by least squares fits and the ends of the series are tapered to zero through the use of a Hanning window. After tapering, complex fast Fourier transforms (FFTs) are computed in longitude and then in time. Finally, the power spectrum is averaged over all segments and over the 15°N-15°S belt. The number of degrees of freedom is calculated to be 81 \[ 2 \text{ (amplitude and phase)} \times 20 \text{ (years)} \times 365 \].
(days)/180 (segment length)]. The signal strength is calculated as $(P_{xx} - P_{red})/P_{xx}$, where

$P_{red}$ is the red spectrum, calculated analytically using Eq. (1) of Masunaga [2007], and values above 0.5 are considered to be statistically significant in this study. We will focus on showing only equatorially symmetric component [Yanai and Murakami, 1970], where the MJO signal is strongest, and to facilitate comparison with other studies. Additionally, the power spectrum is used to calculate the east-west power ratio [Sperber and Kim, 2012], which is defined here as the ratio of intraseasonal eastward vs westward propagating spectral power ($k = 1$-5, 20-100 day timescale).

In the time-longitude lag regression plots shown in Section 5, we estimate the phase speeds and group velocities of the MJO using OLR, precipitation and column water vapor. These calculations are performed following the method described in Adames and Kim [2016]. For the phase speed calculation, we choose extrema (maxima and minima) that occur within 25 days of the reference time (lag day 0). The phase speeds are calculated for each time-longitude section by averaging the MJO-filtered anomaly fields across the longitude intervals ranging from $130^\circ - 145^\circ$E, $145^\circ - 160^\circ$E, $160^\circ - 175^\circ$E, and $175^\circ - 190^\circ$E. For each field, the time when a statistically-significant extremum occurs is calculated within each longitude band. Phase speed is then calculated by linear least squares fit of the time in which an extremum occurs within each longitude band. If multiple propagating envelopes are found, the phase speed is estimated by averaging all of them.

The group velocities are calculated in a similar manner. We calculate the zonal and temporal position of each local extremum. A local extremum is defined here as a local maximum/minimum occurring within 25 days of the reference time. For a local maximum/minimum to be considered, it must be the largest anomaly in space and time within
an interval of 10 days and be significant at the 95\% confidence interval. After all the local
extrema are identified, the group velocity is calculated through a linear least squares fit
in the longitude-time space.

3. Mean state changes with increasing CO$_2$

In order to get an overview of the mean state of the four simulations, the left column
of Fig. 1 shows global distributions of annual mean surface temperatures for the four
GISS simulations. An approximately linear increase in surface temperatures is observed
with each doubling in CO$_2$ concentration. The annual mean column integrated moisture
$\langle q \rangle$ and precipitation $P$ fields are shown in the middle and right columns of Fig. 1,
respectively. A sharp increase in column moisture is evident at all geographical locations,
but in contrast to temperature changes that exhibits the so-called polar-amplification,
moisture increases more over the tropics, consistent with the Clausius-Clapeyron equation.

It is noteworthy that column moisture increases the fastest over the central and eastern
Pacific ITCZ. Unlike moisture, which increases everywhere with warming, not all locations
exhibit increasing precipitation. Instead, precipitation increases the most over the eastern
and central Pacific, consistent with the strongest increments in moisture and consistent
with a weakening of the Walker Circulation with increasing CO$_2$ [Seager et al., 2010]. It
is worth noticing that precipitation decreases over tropical landmasses.

Latitude-height cross sections of the difference in zonally-averaged latent energy ($L_v q$,
where $L_v = 2.5 \times 10^6$ J kg$^{-1}$ is the latent heat of vaporization), dry and moist static
energy and relative humidity between the 0.5CO$_2$ and 4CO$_2$ runs are shown in Fig. 2. An
increase in lower-tropospheric latent energy is observed at all latitudes, with a maximum
near the equator, consistent with Fig. 1. Dry static energy, defined as $s = C_pT + gZ$, where
$C_p = 1004 \text{ J K}^{-1}$ is the specific heat of dry air, increases at all heights and latitudes. Unlike moisture, $s$ has a maximum rate of change near the equator at $\sim 200 \text{ hPa}$, indicating an increase in static stability with warming. Moist static energy (MSE, $m = L v q + s$) increases the most near the surface at the equator, consistent with the sharp increase in specific humidity near the surface. A secondary maximum is observed $\sim 200 \text{ hPa}$ where the dry static energy anomalies are a maximum. Even though latent energy $L v q$ increases the most over the lower troposphere, relative humidity, shown in Fig. 2d increases over the tropical lower free troposphere $\sim 700 \text{ hPa}$ and just below the tropopause $\sim 200 \text{ hPa}$.

In order to quantify the trends in mean moisture and precipitation, scatterplots of column moisture, precipitation and column relative humidity, averaged over the equatorial warm pool ($60^\circ \text{E} - 180^\circ$, $15^\circ \text{N} / \text{S}$ the region is enclosed by dotted lines in Fig. 1), where MJO activity is strongest, are shown in Fig. 3 as a function of the mean tropical temperature. Only points located over the oceans are included. Column moisture increases at a rate of $\sim 7.1\% \text{ K}^{-1}$, similar to that predicted by the Clausius-Clapeyron equation, while precipitation increases at a fraction of this rate ($\sim 3 \% \text{ K}^{-1}$). Global precipitation increases at a smaller rate water vapor because it is highly constrained by the global radiative cooling rate [Allen and Ingram, 2002; Held and Soden, 2006; O'Gorman et al., 2012; Pendergrass and Hartmann, 2014]. The rate of change of warm-pool averaged precipitation is consistent with this notion, augmented by changes in the mean warm pool circulation with increasing CO$_2$ [Chou et al., 2009; Seager et al., 2010]. In contrast, column relative humidity ($R_h = \langle q \rangle / \langle q_s \rangle$), shown in the right panel of Fig. 3, shows little change as temperatures increase over the warm pool.
4. Changes in warm pool variability with increasing CO$_2$

Maps of the standard deviation of daily precipitation and OLR are shown in Fig. 4. Variability in precipitation exhibits peak over the SPCZ in the 0.5CO$_2$ run, while OLR peaks over the western Pacific. As CO$_2$ increases, variability in precipitation amplifies and the peak shifts to the western Pacific near the maximum in OLR variability. A significant amplification in precipitation is also observed over the Indian Ocean and over the ITCZ. In contrast, OLR variability shows a slight decrease in variability. MJO-filtered variability, shown in contours, shows similar trends in both fields.

In order to quantify the changes in variability over the Indo-Pacific warm pool, scatter-plots as in those shown in Fig. 3 but for the root-mean-squared amplitude of anomalous column water vapor $\langle q \rangle$ and precipitation $P$ are shown in Fig. 5. Interestingly, for both precipitation and column water vapor, the changes in variability scales well with those of the mean states (bottom panels of Fig. 5). In other words, the moisture and precipitation variability increases at a same rate with which the climatological mean values increase, suggesting that mean state changes might constrain changes in variability.

In constrast to variability across all temporal and spatial scales, MJO-filtered variability, shown in panels (a) and (b) of Fig. 6, increases at a faster rate than variability across all spatial and temporal scales. A useful quantity that can further show how the MJO amplifies with increasing CO$_2$ is the east-west power ratio, shown in Fig. 6c for precipitation and column water vapor $\langle q \rangle$. These trends are weaker than those of MJO-filtered anomalies, indicating that the amplification occurs across broad intraseasonal bands, not preferentially in the eastward propagating, planetary scale waves (i.e. MJO). This sug-
gests that part of the MJO amplification is due to overall increase in tropical variability, especially in the intraseasonal components.

A more comprehensive understanding of how variability in the model is changing with increasing CO$_2$ can be obtained by performing a space-time spectral analysis over the equatorial belt. Column (1) of Fig. 7 shows the zonal wavenumber-frequency power of column water vapor averaged over the 15°N/S belt. An increase in spectral power is observed at all frequencies and wavenumbers as CO$_2$ increases, consistent with the overall increase in variability shown in Fig. 5. However, the amplification is strongest along the lowest frequencies and wavenumbers, suggesting a systematic “reddening” of tropical water vapor variability with increasing CO$_2$. This reddening is more clearly depicted in the difference in the power spectrum between the 4CO$_2$ and 0.5CO$_2$ simulations, shown in Fig. 7e. This systematic reddening of the column moisture power power spectrum may explain why intraseasonal variations in moisture (Fig. 6a) increase at a faster rate than other spatial and temporal scales, yet the east-west power ratio remains relatively fixed. The spectrum of precipitation (column 5), exhibits a more uniform change in spectral density than water vapor does. A reddening of this spectrum is not as clear as changes are dominated by a strengthening of spectral power along the Kelvin wave dispersion curve.

Columns (2) and (4) of Fig. 7 show the signal strength of water vapor and precipitation, respectively defined as in Hendon and Wheeler [2008] as the amount of signal that lies above a red noise spectrum, $S = (P_{xx} - P_{red})/P_{xx}$. The plots are overlaid by the dispersion curves for Kelvin, n=1 equatorial Rossby, tropical depression-type disturbances as defined by Yasunaga and Mapes [2012], and for the MJO, as defined in Eq. (29a) of Adames and
Kim [2016] (curves are defined in the same way as their Fig. 13). For precipitation, the strongest signal in the MJO band is found between zonal wavenumbers 1-5 in the 0.5CO$_2$ simulation, with some signal occurring at higher wavenumbers. The MJO signal in water vapor is confined within the first three zonal wavenumbers and slightly shifted towards higher frequencies. It is noteworthy that while precipitation exhibits a pronounced Kelvin wave signal, such a signal is largely absent in the column water vapor signal, although an eastward-propagating signal in \(\langle q\rangle\) is seen that is reminiscent of Kelvin waves of shallower equivalent depths. This result is consistent with previous studies that have looked at the spectrum of Kelvin waves [Sobel and Bretherton, 2003; Yasunaga and Mapes, 2012]

As CO$_2$ concentrations increase, the precipitation signal becomes more confined at wavenumbers 1-3 and shifts towards higher frequencies. An increase in frequency is also observed. The difference in signal for both column water vapor and precipitation is shown in the bottom panel of Fig. 7. For both fields, an increase in signal is seen between 20-40 timescale, while a reduction is seen at lower frequencies. This is consistent with the MJO shifting towards higher frequencies in warmer climates (e.g., Arnold et al. 2013).

A strengthening of the Kelvin wave signal in precipitation is observed, combined with a shift towards larger equivalent depth, consistent with the increasing static stability seen in Fig. 2b. However, that no clear Kelvin wave signal is apparent along the water vapor dispersion curves is in agreement with the notion that precipitation in relation to Kelvin wave activity is mainly due to changes in tropospheric temperature with secondary effects from moisture fluctuations [Sobel and Bretherton, 2003; Raymond and Fuchs, 2009; Herman et al., 2016].
Coherence squared and the spectral phase angle between the precipitation and column moisture fields is shown in column (3) of Fig. 7. A strong correspondence is observed between the largest coherence values and the dispersion curves corresponding to the MJO. The high values of coherence along the MJO band are co-located with the MJO-related precipitation signal in column (4) of Fig. 7. Significant coherence is also observed over the Kelvin wave dispersion curve. However, coherence values along the Kelvin wave dispersion curves are much smaller than those seen along the MJO band. As CO$_2$ concentrations increase, the magnitude of the coherence does not significantly change, but becomes more concentrated at the lowest zonal wavenumbers. There is also indication of a shift towards higher frequency, with red shading, corresponding to the largest coherence values, seen near the 20 day timescale band near zonal wavenumber 1. This change is most clearly seen in panel (e), with the difference of coherence between the 0.5 and 4CO$_2$ runs showing increased coherence along zonal wavenumber 1, and decreased coherence along higher zonal wavenumbers. The spectral phase angle, shown as arrows, indicates that column moisture slightly leads the precipitation anomalies, consistent with observations [Yasunaga and Mapes, 2012]. No significant change in the phasing between water vapor and precipitation with warming is observed.

5. Changes in the MJO characteristics with increasing CO$_2$

In order to elucidate the changes in the structure and propagation of the MJO with increasing CO$_2$, a linear regression analysis is performed using a time-varying index. We will focus our analysis on maps based on 20-100 day timescale, eastward propagating wavenumbers 1-5 OLR time series averaged over the western Pacific (140-180°E, 10° N/S), where intraseasonal variability in the GISS simulations is strongest. We have verified that
results shown herein are also quantitatively reproducible using OLR-based time series over
the Maritime Continent or the Indian Ocean, or using principal components from EOF
analysis.

Regression maps of the precipitation field along with the 850 hPa winds is shown in
Fig. 8. The structure of precipitation in the western Pacific can be separated into a sector
of rainfall that is located along the inter-tropical convergence zone (ITCZ) \( \sim 7^\circ \), and a
second maximum centered in the Southern Hemisphere oriented along the south Pacific
convergence zone (SPCZ). As CO\(_2\) concentrations increase, the precipitation anomalies
in both regions amplify. The anomalies also become more zonally extensive, though the
increase in areal extent is most apparent over the ITCZ sector. The wind anomalies also
appear to become more zonally extensive, though no clear change in their amplitude was
found (not shown). This is consistent with Maloney and Xie [2013], who suggested that
the tropical dry static stability is as important as rainfall in determining horizontal wind
variability.

Longitude-height cross sections of the specific humidity field, and its local temporal
tendency \( \partial q'/\partial t \) are shown in Fig. 9. An amplification of the specific humidity anomalies
is clearly evident. Additionally, a deepening of the moisture anomalies is also observed,
with the height of the deepest shading in each panel occurring higher in the troposphere
as CO\(_2\) concentrations increase. The location of the regions of maximum moistening to
the east, and drying to the west are similar to those found by Chikira [2014]; Adames and
Wallace [2015] and Wolding and Maloney [2015]. The moisture tendency field also shows
hints of deepening as CO\(_2\) concentrations increase.
Time-longitude diagrams of $\langle q' \rangle$, precipitation and OLR are shown in Fig. 10. These are lag regressions constructed using the OLR time series for the western Pacific sector. Maxima in the three fields (depicted as circles in Fig. 10) become progressively more separated from each other, indicative of an increase in the MJO's group velocity. As a result of these changes, the MJO-related anomalies propagate over a larger area of the tropics in a warmer climate. Whereas it propagates from $\sim$120-200$^\circ$ of longitude in the 0.5CO$_2$ simulation, it propagates from $\sim$60-220$^\circ$ in the 4CO$_2$ simulation. These results are generally consistent with studies by Arnold et al. [2013, 2015]; Chang et al. [2015] among others, who also found the MJO to become faster and cover a larger area of the globe as the climate warms.

Quantitative estimates of the changes in MJO shown in Figs. 8 - 10 are shown in Fig. 11. Fig. 11a shows the mean wavenumber obtained from spatial spectral analysis of the three fields shown in Fig. 10. The decreasing trend seen in Fig. 11a is consistent with the larger zonal extent of the moisture and precipitation anomalies seen in Figs. 8 - 10. The scatterplot indicates that the MJO-related moisture and precipitation anomalies are increasing in zonal extent at a rate of approximately $\sim$150 km per degree of tropical warming, and the zonal scale of the anomalies in the 4CO$_2$ simulation is roughly 500 km larger in zonal extent than the anomalies in the 0.5CO$_2$ simulation. It is noteworthy the MJO's zonal wavenumber in all four GISS simulations is still larger than observations, which was estimated at 1.81 by Adames and Kim [2016]. Chikira and Sugiyama [2013] also found that their simulated MJO was smaller in horizontal extent than the observed MJO.
The region of maximum tropospheric moisture anomalies, shown in Fig. 11b, reveals a steady increase in the height of the moisture maximum. Finally, estimates of the rate of change of the phase speed and group velocity estimates from Fig. 10 are shown in Fig. 11c-d. The phase speed and group velocity of the MJO increase at similar rates of $\sim 3.3\% K^{-1}$ and $\sim 2.6\% K^{-1}$, respectively. A quantitative analysis of the changes in the MJO’s phase speed and group velocity is presented in part II of this study.

6. Concluding Discussion

In this study, we investigated the changes in the MJO as CO$_2$ concentrations increase. Four 20-year long simulations from the NASA-GISS model with a mixed layer ocean (Q-flux), and a modified convection scheme [Del Genio et al., 2012], with CO$_2$ concentrations ranging from half to quadruple of preindustrial levels are analyzed. The model exhibits MJO-like variability reminiscent to observations. This study adds to previous studies on how the MJO changes with increasing CO$_2$ [Arnold et al. 2015; Chang et al. 2015; among others].

Results from the four simulations support many previously documented changes in the mean state associated with greenhouse gas induced warming. Column integrated water vapor concentrations increase following the Clausius-Clapeyron equation ($\sim 7\% K^{-1}$), while mean precipitation increases at a fraction of that rate ($\sim 3\% K^{-1}$) and column relative humidity remains approximately fixed. Furthermore, zonal cross sections reveal increasing relative humidity in the upper-troposphere, increasing static stability and a larger vertical moisture gradient. Moreover, variability over the Indo-Pacific warm pool ($60^\circ E - 180^\circ$, $15^\circ N/S$), defined as the domain-averaged standard deviation, increases at nearly the same rate as the mean state does. However, it is found that moisture and
precipitation variability over the MJO’s spatial and temporal scales (20-100 day timescale, eastward-propagating zonal wavenumbers 1-5) increases at a faster rate of $\sim 9 \, \% \, K^{-1}$ and $\sim 5.6 \, \% \, K^{-1}$, consistent with the results of Arnold et al. [2013, 2015].

Does the amplification of MJO variability (i.e. power) indicate a destabilization of the wave with warming, or is it a consequence of the overall tropical rain variability change? To address this question, the east-west power ratio, the ratio between spectral power over the MJO band and its westward counterpart, is obtained from each simulation and its relationship with warming is examined. If the amplification is a result of a stronger destabilization of the MJO in a warmer climate, the ratio will show strong positive trend as the MJO variability shows. The east-west power increases only modestly with warming at a rate that is much smaller than that for the MJO variability in both precipitation ($\sim 1.8 \, \% \, K^{-1}$) and moisture ($\sim 0.6 \, \% \, K^{-1}$), suggesting that it is likely that the increase in MJO variability with warming is mainly associated with changes in tropical rainfall variability. Therefore, one should be cautious when linking the amplification of intraseasonal anomalies to changes in MJO maintenance (i.e. destabilization).

Through the use of space-time spectral anomalies and linear regression analysis of OLR over the western Pacific, the changes in the MJO as surface temperatures increase were documented. Results from the spectral analysis reveal a shift towards lower wavenumbers (larger scale) and higher frequency (faster propagation) in the MJO, consistent with previous studies [Arnold et al., 2013, 2015; Chang et al., 2015]. This change is evident in the precipitation and moisture signal, as well as their coherence squared.

Linear regression analysis further reveals that the increase in the MJO phase speed is at a rate of $\sim 3.3 \, \% \, K^{-1}$. Consistent with the study of Adames and Kim [2016], a westward
group velocity of the MJO was also found in this study, whose amplitude increases with surface temperatures at a rate of $\sim 2.6\% \text{K}^{-1}$. It is also found that the MJO-related moisture anomalies become deeper as CO$_2$ concentrations increase, with the peak moisture anomalies occurring $\sim 13 \text{hPa}$ higher in the troposphere per degree of tropical warming. This deepening of the moisture anomalies may be due to the deepening troposphere that results from warming, along with changes in the vertical profile of temperature, which affects moisture through the Clausius-Clapeyron equation.

While the results shown in this study give us some insight onto the trends in the structure and propagation of the MJO with climate change, they do not provide sufficient physical and quantitative insights as to why these changes occur in the GISS simulations. A quantitative analysis of the changes in the propagation and maintenance of the MJO-related moisture anomalies is presented in Part II.

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References


Figure 1. (Left) Mean November-April (NDJFMA) surface temperature, (middle) column water vapor and (right) precipitation for the (a) 0.5CO$_2$, (b) 1CO$_2$, (c) 2CO$_2$ and (d) 4CO$_2$ GISS model simulations. The difference between the 4CO$_2$ and 0.5CO$_2$ simulations is shown in panel (e).
Figure 2. Differences between the 4CO$_2$ and 0.5CO$_2$ simulation NDJFMA zonally averaged (a) latent energy $L_vq$, (b) dry static energy $s = C_p T + gZ$, (c) moist static energy $m = L_vq + s$, and (d) relative humidity $q/q_s$. 
Figure 3. Scatterplots showing NDJFMA mean, warm pool averaged (60°E-180°, 10°N/S) (a) column water vapor $\langle q \rangle$, (b) precipitation and (c) column relative humidity as a function of the mean surface temperature averaged over the 30°N/S latitude belt. The dashed line in each panel corresponds to the nonlinear least squares fit of the trend in each variable.
Figure 4. Standard deviation of the NDJFMA anomalous precipitation (shaded left) and MJO filtered (contoured left) precipitation, and anomalous OLR (shaded right) and and MJO filtered OLR (contoured right) for the (a) 0.5CO$_2$, (b) 1CO$_2$, (c) 2CO$_2$ and (d) 4CO$_2$ simulations.
Figure 5. Scatterplots showing NDJFMA standard deviation of column water vapor $\langle q \rangle$ (left) and precipitation (right) as a function of tropical surface temperature (top) and NDJFMA-mean moisture and precipitation (bottom). The dashed line in each panel corresponds to the nonlinear least squares fit of the trend in each variable. The percentage change of the anomalies per degree of warming is shown in the top panels, and the correlation coefficient between the NDJFMA standard deviation and mean is shown in the bottom panels.
Figure 6. Scatterplots showing NDJFMA, MJO-filtered standard deviation of (a) column water vapor $\langle q \rangle$ and (b) precipitation as a function of tropical surface temperature. (c) East-west power ratio of precipitation (circles) and column water vapor (triangles) for the four GISS simulations. The east-west power ratio is calculated as the ratio in spectral power between eastward-propagating wavenumber averaged over zonal wavenumbers 1-5 and 20-100 day timescales. The percentage-based rate of changes are shown in the figure. The dashed line in each panel corresponds to the nonlinear least squares fit of the trend in each variable.
Figure 7. Space-time spectral analysis of column water vapor \((q)\) and precipitation \(P\): (1 and 5) Normalized power spectrum of the symmetric component of \((q)\) and \(P\), respectively, over the 15°S–15°N latitude band. (2 and 4) Signal strength of \((q)\) and precipitation \(P\), respectively. (3) Coherence squared (shading) and phase angle (arrow) between \((q)\) and precipitation \(P\). Upward-pointing vector corresponds to a zero phase lag, downward implies out of phase, rightward implies that \((q)\) leads \(P\) by a quarter cycle, and leftward implies \((q)\) lags \(P\) by a quarter cycle. The rows show the spectral analysis corresponding to the (a) 0.5CO2, (b) 1CO2, (c) 2CO2 and (d) 4CO2 simulations and (e) the difference between the 4CO2 and 0.5CO2 simulations. Dispersion curves are plotted in columns (2)-(4) for Kelvin and \(n=1\) equatorial Rossby waves, with equivalent depths of 12, 25 50, and 90 m, respectively. Dotted lines indicate constant phase speeds of 7.0, 9.0, and 11.0 m s\(^{-1}\), which are representative of westward-propagating tropical depression and easterly waves (see also Yasunaga and Mapes 2012). Contour interval is every 0.05 signal strength fraction beginning at 0.5. The solid lines correspond to MJO dispersion curves as derived in Eq. (29a) in Adames and Kim [2016].
Figure 8. Linear regression for precipitation (shaded) and 850 hPa wind anomalies (arrows) onto an MJO-filtered time series centered over the western Pacific. Each panel corresponds to the anomaly fields for the (a) 0.5CO$_2$, (b) 1CO$_2$, (c) 2CO$_2$ and (d) 4CO$_2$ simulations. The largest arrows correspond to wind anomalies of $\sim$ 1 m s$^{-1}$.
Figure 9. Composite longitude-height cross section of latent energy anomalies $L_v q'$ (shaded), its temporal tendency $L_v \partial q'/\partial t$ and the anomalous zonal mass circulation $(\rho u', \rho w')$ for the (a) 0.5CO$_2$, (b) 1CO$_2$, (c) 2CO$_2$ and (d) 4CO$_2$ simulations. The largest zonal flux vector is $\sim 0.2$ kg m$^{-2}$ s$^{-1}$. Contour interval 30 J kg$^{-1}$ day$^{-1}$. 
Figure 10. Time-longitude diagrams of 20-100 day timescale filtered (shaded) and MJO filtered (contours) \(q\) (left column), precipitation (middle column) and OLR (right column) for the (a) 0.5CO\(_2\), (b) 1CO\(_2\), (c) 2CO\(_2\) and (d) 4CO\(_2\) simulations. The reference time corresponds to the time when the MJO-filtered anomalies are a minimum over the western Pacific. The gray dashed lines are linear least squares fit estimates of the phase speed and group velocity for the MJO filtered fields. Circles correspond to the local extremum of each field. Estimate phase speed and group velocities, along with their uncertainties, are shown in the top-left corner of each diagram. Contour interval 1 Wm\(^{-2}\) for OLR and 0.25 mm for \(q\)'.
Figure 11. Statistics of the composite MJO for the four GISS simulations. For the mean zonal wavenumber, a spectral analysis in longitude of the anomalies in Fig. 10 is performed, for all days within 25 days of the reference time. The power spectrum is then averaged for all the latitudes and days included and then normalized using the formula $\tilde{P}_{xx}(k) = P_{xx}(k)/\Sigma_{k=1}^{N} P_{xx}(k)$. The approximate wavenumber $k$ is obtained by summing the zonal wavenumbers, weighting each one by its normalized power $\tilde{P}_{xx}$. The phase speed and group velocities are averaged from the phase speed and group velocities in Fig. 10. The dashed line in each panel corresponds to the nonlinear least squares fit of the trend in each variable.