Circulation, Moisture, and Precipitation Relationships along the South Pacific Convergence Zone in Reanalyses and CMIP5 Models

MATTIE NIZNIK AND BENJAMIN R. LINTNER

Department of Environmental Sciences, Rutgers, The State University of New Jersey, New Brunswick, New Jersey

(Manuscript received 2 May 2013, in final form 30 July 2013)

ABSTRACT

One theorized control on the position of the South Pacific convergence zone (SPCZ) is the amount of low-level inflow from the relatively dry southeastern Pacific basin. Building on an analysis of observed SPCZ region synoptic-scale variability by Lintner and Neelin, composite analysis is performed here on two reanalysis products as well as output from 17 models in phase 5 of the Coupled Model Intercomparison Project (CMIP5). Using low-level zonal wind as a compositing index, it is shown that the CMIP5 ensemble mean, as well as many of the individual models, captures patterns of wind, specific humidity, and precipitation anomalies resembling those obtained for reanalysis fields between weak- and strong-inflow phases. Lead–lag analysis of both the reanalyses and models is used to develop a conceptual model for the formation of each composite phase. This analysis indicates that an equatorward-displaced Southern Hemisphere storm track and an eastward-displaced equatorial eastern Pacific westerly (wind) duct are features of the weak-inflow phase although, as indicated by additional composite analyses based on these features, each appears to account weakly for the details of the low-level inflow composite anomalies. Despite the presence of well-known biases in the CMIP5 simulations of the SPCZ region climate, the models appear to have some fidelity in simulating synoptic-scale relationships between low-level winds, moisture, and precipitation, consistent with observations and simple theoretical understanding of interactions of dry air inflow with deep convection.

1. Introduction

The most prominent feature of South Pacific climate is the South Pacific convergence zone (SPCZ), an area of low-level convergence and strong convective precipitation. Bergeron (1930) and Hubert (1961) were the first to identify the SPCZ in surface observations and early satellite cloud imagery, respectively, while the term SPCZ is attributed to Trenberth (1976) (Kiladis et al. 1989; Vincent 1994; Brown et al. 2011b). The SPCZ climatological location spans tropical and subtropical latitudes, extending from the equatorial western Pacific warm pool in the northwest to Southern Hemisphere midlatitudes near ~30°S, 120°W (Trenberth 1976; Kiladis et al. 1989; Vincent 1994). The SPCZ is stronger in austral summer than in other seasons (Meehl 1987; Vincent 1994), during which time it is responsible for a large fraction of precipitation occurring across the South Pacific (Brown et al. 2011b). Consequently, the SPCZ is of relevance to agriculture, food security, and human health sectors for many island nations across the South Pacific and, as such, has been an important focus of studies of both present-day climate [e.g. Southwest Pacific Ocean Circulation and Climate Experiment (SPICE), see Ganachaud et al. (2007)] and projected future climate [e.g. the Pacific–Australia Climate Change Science and Adaptation Planning (PACCSAP) program, see Australian Bureau of Meteorology and CSIRO (2011a,b)] in the South Pacific. However, current understanding of the SPCZ remains incomplete: Power (2011) identified the need for additional effort “to increase understanding [of the SPCZ] on many fronts, including the reasons why the SPCZ exists.”

The location and intensity of SPCZ precipitation are highly variable and are influenced by multiple factors operating on different time scales. Power spectra analyses of SPCZ variability, focusing on time scales less than 90 days, reveal that the more zonal tropical SPCZ, roughly defined as the portion north of 20°S, shows strong peaks at two weeks and one to two months, while the subtropical SPCZ, lying south of 20°S, varies strongly on time scales of one week or less (Widlansky et al. 2011; Matthews 2012). The synoptic variability of the subtropical...
SPCZ as revealed through these analyses validates similar findings using early satellite imagery (Streten 1973). The tropical SPCZ results suggest longer synoptic (two week) variability in this region in addition to interactions with the Madden–Julian oscillation (MJO) on intraseasonal (30–60 day) time scales (Matthews et al. 1996; Matthews 2012). Beyond seasonal time scales, the SPCZ interacts with the El Niño–Southern Oscillation (ENSO) on interannual time scales (Trenberth 1976; Folland et al. 2002; Vincent et al. 2011), as well as the interdecadal Pacific Oscillation (IPO) on decadal time scales (Folland et al. 2002; Linsley et al. 2008). Particularly strong El Niños may trigger so-called zonal SPCZ events in which the SPCZ effectively collapses onto the equator and merges with the intertropical convergence zone (ITCZ), thereby altering the hydroclimate and its extremes across the South Pacific (Cai et al. 2012). The interannual modulation of the position of the SPCZ in turn affects source regions of South Pacific tropical cyclogenesis (Vincent et al. 2011).

Several mechanisms have been posited to explain the origin and variability of the SPCZ. The extent of the Andes-forced subsidence, or “dry zone,” over the southeastern tropical Pacific is thought to limit the eastward extent of the SPCZ (Takahashi and Battisti 2007). Inhibition of precipitating deep convection within the dry zone establishes a climatological background state upon which the SPCZ varies, including the SPCZ’s characteristic northwest-to-southeast axis (or “diagonal” orientation). High temporal frequency (∼synoptic) changes in low-level inflow from the dry zone are related to shifts in the convective margin on the eastern flank of the SPCZ, with increased low-level inflow (e.g., stronger trade winds) shifting the convective margin to the west (Lintner and NeeLin 2008, hereafter LN08). The SPCZ region has also been described in terms of a “storm graveyard” (Trenberth 1976), where upper-level negative zonal stretching deformation (i.e., \( \partial U/\partial x < 0 \)) slows the group speed of eastward-propagating synoptic disturbances, thereby decreasing wavenumber and increasing wave energy density (Widlansky et al. 2011). This forcing can in turn trigger deep convection, due to the high sea surface temperatures (SSTs) and conditional instability in the SPCZ region (Matthews 2012). Additionally, the first two empirical orthogonal functions (EOFs) of outgoing longwave radiation (OLR) in the region 10°–15°S, 175°E–180° (within the tropical SPCZ) are 1) a shift in the position perpendicular to its axis of maximum precipitation and 2) “pulses” of energy that enhance SPCZ precipitation; accordingly, the climatological SPCZ can be viewed as the sum of these pulse events (Matthews 2012).

In terms of climate modeling, members of both the World Climate Research Program (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) (Meehl et al. 2007) and phase 5 (CMIP5) (Taylor et al. 2012) show limited success in simulating the SPCZ region. Previous analyses of GCM-simulated SPCZs have tended to show overly zonal orientations (Brown et al. 2011b; J. R. Brown et al. 2013), with many models still apparently simulating one ITCZ in each hemisphere, that is, the well-known double ITCZ bias (Zhang 2001; Lin 2007; de Szeoke and Xie 2008; Bellucci et al. 2010; Brown et al. 2011b; J. R. Brown et al. 2013). The double ITCZ bias may not persist throughout the year; some models simulate an ITCZ that shifts toward the summer hemisphere, thus leaving the appearance of two ITCZs in the annual mean, while others confine this bias to the eastern Pacific (de Szeoke and Xie 2008; Bellucci et al. 2010; Brown et al. 2011b). Both the double ITCZ bias and “zonal” bias have been tied to errors in SST across the South Pacific with cooler than observed equatorial SSTs, the “cold tongue,” playing a key role (Ashfaq et al. 2010; Widlansky et al. 2013). In turn, these biases alter the climatological position of the storm graveyard (Widlansky et al. 2011). Output from the CMIP5 Atmospheric Model Intercomparison Project (AMIP) (Taylor et al. 2012) experiments confirms that the biases are reduced when observed SSTs are used, though convergence zone precipitation remains more intense than observed (Widlansky et al. 2013).

These pathological errors in SPCZ simulation lead to lowered confidence in projected anthropogenic warming-induced changes across the South Pacific, which may include alterations to the spatial distribution and magnitude of rainfall, a shift in the SPCZ’s maximum precipitation axis, and increased frequency of zonal SPCZ events (Brown et al. 2011a; J. R. Brown et al. 2013; Widlansky et al. 2013; Cai et al. 2012). For example, while CMIP5 models broadly simulate an increase in SPCZ precipitation (except along its eastern margin), models forced with bias-adjusted SST changes suggest that a drying trend in the SPCZ is possible until tropical SSTs rise above a given threshold of about 3°C (Brown et al. 2011a; Widlansky et al. 2013). Given these issues with the mean SPCZ state as well as coupled model SST biases, there is no guarantee that synoptic-scale SPCZ variability is well represented in GCMs. In fact, LN08 suggest that errors in the way models simulate high-frequency interactions of low-level circulation, moisture, and precipitation could contribute to SPCZ biases. Thus, examining synoptic-scale variability in these models may provide insight into their climatological biases, stimulate model improvements, and increase confidence in future projections (J. N. Brown et al. 2013).

It is difficult to make any supposition about the ability of coupled GCMs to capture any synoptic-scale SPCZ
variability, as it has not been the focus of any recent studies. The primary purpose of this work is to examine one source of synoptic-scale SPCZ variability, changes to low-level inflow east of the SPCZ, as simulated by CMIP5 models. Comparing such variability to available reanalysis products may help to elucidate why current generation GCMs have difficulties simulating the SPCZ and, more broadly, the South Pacific. The rest of this paper is organized as follows. Section 2 outlines the data and analysis methodology used in this paper. Section 3 is primarily concerned with the examination of the new daily composite analysis using reanalysis products [National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (R1) (Kalnay et al. 1996) and the NCEP Climate Forecast System Reanalysis (CFSR) (Saha et al. 2010); see section 3a] and GCMs (section 3b) at the compositing level, as well as the vertical extent and consistency of these anomalies in both the reanalysis products and the GCMs (section 3c). Section 4 contains a description of the results of the lead–lag analysis (composite plots of multiple variables in the days preceding and following the extrema of the composite index) for both the reanalysis and models. Finally, a summary of these findings and their implications are provided in section 5.

2. Data and methodology

LN08 used 5-day- (pentadal-) averaged 925-hPa zonal and meridional winds and 850-hPa specific humidity data at 2.5° × 2.5° resolution derived from daily average output from the R1 reanalysis (Kalnay et al. 1996) in their composite plots of the difference in wind, moisture, and precipitation across the South Pacific between the positive and negative phase of their composite index. These phases were defined by the strength of zonal wind in the box 20°–10°S, 140°–120°W: the positive and negative phases contained all events for which the value was +1 σ and −1 σ, respectively (see LN08 for a more detailed description). Although the use of pentadal data suppresses noise, it precludes exploration of the growth and decay of observed anomalies on daily time scales. Based on preliminary analysis of model and reanalysis data showing little sensitivity to the choice of daily data over pentadal data (further confirmed by the similarity between pentadal and daily products in LN08), all analyses in this paper are performed using daily averages. We use R1 data as a starting point, but further examine daily averages in the same fields and resolution from the CFSR (Saha et al. 2010) during the 32-yr period spanning 1979–2010 (available for both reanalyses). In addition to having approximately six times the resolution in the horizontal and double the number of vertical levels, the CFSR also improves upon R1 by including a coupled ocean model and updating many of the physical parameterizations in the atmospheric model (see Saha et al. 2010 for a detailed discussion). Figure 1 illustrates R1 and CFSR January (chosen for consistency with LN08 and computational efficiency) precipitation climatologies across the South Pacific for 1998–2010, the subset of years for which Tropical Rainfall Measuring Mission (TRMM) precipitation estimates (specifically, the 3B42 dataset) are available (Kummerow et al. 2000). While CFSR output matches TRMM estimates fairly well, R1 precipitation is too zonal, consistent with model biases evident in both CMIP3 and CMIP5 model suites (Brown et al. 2011b; J. R. Brown et al. 2013). While there are potentially significant issues regarding the reliability of reanalyzed specific humidity and precipitation (e.g., model biases in regions with sparse observations), we suggest that the more realistic SPCZ in CFSR is a better
reference moving forward until an appropriate observational dataset of equal length is available at daily time scale.

Table 1 lists the 26 models for which output is examined in this work; all are included in CMIP5 (Taylor et al. 2012). These models were selected based on the availability of daily precipitation as well as pressure- and wind, meridional wind, and specific humidity for at least one ensemble member included in the CMIP5 historical experiment (spanning 1850–2005). This experiment includes both observed anthropogenic and natural forcings during that period (Taylor et al. 2012).

We further examine the output of these models (with one exception) from the high emissions RCP8.5 projection spanning 2006–2300 (Taylor et al. 2012). This scenario, the strongest of the forcing scenarios in CMIP5, is similar to the choice of the A2 scenario in Brown et al. (2011a) for diagnosing future SPCZ change in CMIP3 models, though it should be noted that Widlansky et al. (2013) show the SPCZ response to warming is likely nonlinear. With a few exceptions (see Table 1), the first ensemble member for the CMIP5 historical and RCP8.5 experiments at daily resolution was obtained and regridded from the resolutions shown in Table 1 onto a 2.5° × 2.5° common grid, which is chosen because it matches the resolution of the data used in LN08. Models in which the product of native grid latitude and longitude is less than 6.25° squared were regridded to the analysis domain via area averaging; the remainder were regridded using linear interpolation.

The analysis intervals are 1960–99 and 2060–99 for the historical experiment and RCP8.5 experiment, respectively. Selection of these intervals was motivated by the greater availability of model output in these periods and the desire to have two periods of equal duration. The first 30 days of January were isolated in each year, yielding a total of 1200 days for both the historical data and RCP8.5 data. Any models lacking zonal wind, meridional wind, specific humidity, and/or precipitation anomalies; wind vectors are considered significant if either the zonal or meridional wind anomaly exceeds 99% of these random departures from the weak-inflow composite. The weak-inflow phase (occurring when low-level inflow is weak, previously “negative phase” in LN08) was calculated with the initial day of December 1960 and the final day of November 1999, yielding a total of 153 days for the reanalysis periods, 192 days for the model periods), consistent with the percentage of data expected to lie in either the positive or negative tail of a normal distribution with magnitudes above ±1σ. Moisture and precipitation anomalies are considered significant if they are greater than 99% of these random anomalies; wind vectors are considered significant if either the zonal or meridional wind anomaly exceeds 99% of the

An important difference from the LN08 analysis is our definition of low-level inflow. While LN08 based their composites on zonal wind at 925 hPa, we select 850 hPa as the composite level as well as the lowest plotting level for wind since the PCMDI-archived CMIP5 output does not include the 925-hPa level at daily resolution. In prior analyses with CMIP3 data, which have zonal and meridional winds at both 925 and 850 hPa at daily resolution, we examined the sensitivity of our results to the use of either 925 or 850 hPa and found negligible change. This is consistent with the findings of LN08. In fact, this vertical consistency is not limited to these two levels and will be discussed further in section 3c.

For each day, the areal-mean zonal wind at 850 hPa (mean $u_{850}$) was calculated within the region 20°–10°S, 140°–120°W. The position of the region is unadjusted for inclusion in the weak-inflow composite, the index exceeds a threshold relative to the location of the SPCZ in each model, as the results are relatively insensitive to the exact longitude of the box. These values were first normalized by subtracting the January-mean $u_{850}$ for each year in order to remove any biases due to low frequency variability such as the El Niño–Southern Oscillation (ENSO). Thus, our composite index is centered around zero and measures mean $u_{850}$ departures from the yearly (January) mean. Previously, LN08 had excluded those Januaries in which a strong El Niño or La Niña occurred; this work does not since the results are insensitive to this exclusion. The weak-inflow phase (occurring when low-level inflow is weak, previously “positive phase” in LN08) and strong-inflow phase (occurring when low-level inflow is strong, previously “negative phase” in LN08) composites of zonal and meridional wind, specific humidity, and precipitation are then calculated by averaging all days for which the composite index exceeds a threshold relative to the standard deviation of the composite index ($\sigma$); that is, for inclusion in the weak-inflow composite, the index must exceed $+1\sigma$ while for the strong-inflow composite, it must be less than $-1\sigma$. Composite difference plots are then generated by subtracting the strong-inflow composite from the weak-inflow composite.

To quantify the significance of the wind, moisture, and precipitation anomalies, 1000 random weak-inflow and strong-inflow phases were generated for each reanalysis product and model, each containing approximately 16% of available days ($-153$ days for the reanalysis periods, 192 days for the model periods), consistent with the percentage of data expected to lie in either the positive or negative tail of a normal distribution with magnitudes above $1\sigma$. Moisture and precipitation anomalies are considered significant if they are greater than 99% of these random anomalies; wind vectors are considered significant if either the zonal or meridional wind anomaly exceeds 99% of the
TABLE 1. CMIP5 models used in this paper. All models had daily data available from CMIP5 historical experiment runs and are used in the contemporary period (1960–99) analysis. Those models in italics are not used in the future period (2060–2100) analysis; the remaining 17 are members of the model subset. Except for FGOALS-g2 and CCSM4 (for which the third and sixth ensemble members, respectively, were used due to availability), the first ensemble member of the historical and RCP8.5 experiment was chosen. All models were regridded to 2.5° in both latitude and longitude using area averaging (if in the subset) or linear interpolation (if not in the subset*) for purposes of comparison and the calculation of a model ensemble mean. Latitude and longitude columns list the resolution of the model output available from the PCMDI CMIP5 database. Further information can be found online (http://cmip-pcmdi.llnl.gov/cmip5/docs/CMIP5_modeling_groups.pdf).

<table>
<thead>
<tr>
<th>Modeling group</th>
<th>Model name</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beijing Climate Center (BCC), China Meteorological Administration</td>
<td>BCC Climate System Model, version 1.1 (BCC-CSM1.1)</td>
<td>2.81</td>
<td>2.81</td>
</tr>
<tr>
<td>College of Global Change and Earth System Science, Beijing Normal University (BNU)</td>
<td>BNU - Earth System Model (BNU-ESM)</td>
<td>2.81</td>
<td>2.81</td>
</tr>
<tr>
<td>Canadian Centre for Climate Modeling and Analysis</td>
<td>Second Generation Canadian Earth System Model (CanESM2)</td>
<td>2.81</td>
<td>2.81</td>
</tr>
<tr>
<td>National Center for Atmospheric Research (NCAR)</td>
<td>Community Climate System Model, version 4 (CCSM4, r6)</td>
<td>0.94</td>
<td>1.25</td>
</tr>
<tr>
<td>Centro Euro-Mediterraneo per I Cambiamenti Climatici (CMCC)</td>
<td>CMCC Carbon Earth System Model (CMCC-CESM)</td>
<td>3.75</td>
<td>3.75</td>
</tr>
<tr>
<td></td>
<td>CMCC Climate Model (CMCC-CM)</td>
<td>0.75</td>
<td>0.75</td>
</tr>
<tr>
<td></td>
<td>CMCC Climate Model with a Resolved Stratosphere (CMCC-CMS)</td>
<td>1.88</td>
<td>1.88</td>
</tr>
<tr>
<td>Centre National de Recherches Meteorologiques (CNRM)/Centre Europeen de Recherche et Formation Avancees en Calcul Scientifique</td>
<td>CNRM Coupled Global Climate Model, version 5 (CNRM-CM5)</td>
<td>1.41</td>
<td>1.41</td>
</tr>
<tr>
<td>Commonwealth Scientific and Industrial Research Organisation (CSIRO) in collaboration with the Queensland Climate Change Centre of Excellence</td>
<td>CSIRO Mark, version 3.6.0 (CSIRO-Mk3.6.0)</td>
<td>1.88</td>
<td>1.88</td>
</tr>
<tr>
<td>Key Laboratory of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, and Center for Earth System Science (CESS), Tsinghua University</td>
<td>Flexible Global Ocean–Atmosphere–Land System Model gridpoint, version 2 (FGOALS-g2, r3)</td>
<td>3.00</td>
<td>2.81</td>
</tr>
<tr>
<td>LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences</td>
<td>Flexible Global Ocean–Atmosphere–Land System Model gridpoint, second spectral version (FGOALS-s2)</td>
<td>1.67</td>
<td>2.81</td>
</tr>
<tr>
<td>National Oceanic and Atmospheric Administration/Geophysical Fluid Dynamics Laboratory</td>
<td>Geophysical Fluid Dynamics Laboratory Climate Model, version 3 (GFDL-CM3)</td>
<td>2.00</td>
<td>2.50</td>
</tr>
<tr>
<td></td>
<td>Geophysical Fluid Dynamics Laboratory Earth System Model with Generalized Ocean Layer Dynamics (GOLD) component (GFDL-ESM2G)</td>
<td>2.00</td>
<td>2.50</td>
</tr>
<tr>
<td></td>
<td>Geophysical Fluid Dynamics Laboratory Earth System Model with Modular Ocean Model 4 (MOM4) component (GFDL-ESM2M)</td>
<td>2.00</td>
<td>2.50</td>
</tr>
<tr>
<td>Met Office Hadley Centre</td>
<td>Hadley Centre Global Environment Model, version 2 - Carbon Cycle (HadGEM2-CC)</td>
<td>1.25</td>
<td>1.88</td>
</tr>
<tr>
<td>Institute for Numerical Mathematics (INM)</td>
<td>INM Coupled Model, version 4.0 (INM-CM4.0)</td>
<td>1.50</td>
<td>2.00</td>
</tr>
<tr>
<td>Institut Pierre-Simon Laplace (IPSL)</td>
<td>IPSL Coupled Model, version 5, coupled with NEMO, low resolution (IPSL-CM5A-LR)</td>
<td>1.88</td>
<td>3.75</td>
</tr>
<tr>
<td></td>
<td>IPSL Coupled Model, version 5, coupled with NEMO, mid resolution (IPSL-CM5A-MR)</td>
<td>1.26</td>
<td>2.50</td>
</tr>
<tr>
<td>Atmosphere and Ocean Research Institute (University of Tokyo), National Institute for Environmental Studies, and Japan Agency for Marine–Earth Science and Technology</td>
<td>Model for Interdisciplinary Research on Climate, version 4 (high resolution) (MIROC4h)*</td>
<td>0.56</td>
<td>0.56</td>
</tr>
<tr>
<td></td>
<td>Model for Interdisciplinary Research on Climate, version 5 (MIROC5)</td>
<td>1.41</td>
<td>1.41</td>
</tr>
<tr>
<td>Japan Agency for Marine–Earth Science and Technology, Atmosphere and Ocean Research Institute (University of Tokyo), and National Institute for Environmental Studies</td>
<td>Model for Interdisciplinary Research on Climate, Earth System Model, Chemistry Coupled (MIROC-ESM-CHEM)</td>
<td>2.81</td>
<td>2.81</td>
</tr>
<tr>
<td></td>
<td>Model for Interdisciplinary Research on Climate, Earth System Model (MIROC-ESM)</td>
<td>2.81</td>
<td>2.81</td>
</tr>
</tbody>
</table>
random anomalies in the same direction. Similarly, for plots that show anomalies in the weak- and strong-inflow phases compared to the mean state, anomalies are considered significant if they are greater than 99% of these random differences from the mean state.

Lead–lag composites at “day 0” were generated using the same composite index, but restricted to only those days defined as the local maxima and minima during weak-inflow and strong-inflow phase events, respectively. Here, maxima or minima are defined such that a given day meets the standard deviation threshold for inclusion in the previous weak-inflow (strong inflow) composite plots and has an index value greater than (less than) the two preceding and two following days. (Since the index is only calculated in January, the first two days of each month lack two preceding days and are thus never considered as peaks; likewise, the last two days of each month lack two following days and are also excluded.) This methodology was chosen to ensure that certain anomalies, such as synoptic disturbances, are not completely lost in averaging since weak-inflow and strong-inflow phase events can occur across successive days. The included composite plots, which subtract the mean January state of the reanalysis/model from the weak-inflow or strong-inflow composite, allow adequate determination of those features associated with the rise and decay of each phase of the composite index. To facilitate identification of anomalous cyclones and anticyclones, sea level pressure (SLP) is also analyzed here.

3. Low-level inflow composites

a. Reanalyses

Figure 2 depicts composite differences for (a) 850-hPa winds and moisture and (b) precipitation for R1 (left) and CFSR (right). Note that the R1 results are quite similar to those in LN08 using pentadal data for wind and moisture (cf. their Figs. 1a and 1b). In addition, application of a significance test confirms that the wind and moisture anomalies near the compositing region, that is, the northern side of the anomalous cyclonic circulation centered at 30°S, 130°W, are significant. The R1 precipitation field reveals an interesting, albeit not completely unexpected, divergence from the previously used CMAP fields of LN08: instead of showing a shift in convection along the southeastern edge of the SPCZ associated with weakened easterlies in the compositing region, R1 indicates a simple expansion of precipitation farther to the southeast, thereby (partially) overcoming its tendency to simulate an overly zonal and tropical SPCZ. This result is consistent with many of the CMIP5 models that exhibit a similar climatological SPCZ orientation (see section 3b). Additionally, the significance of the precipitation expansion is now confirmed.

The CFSR results compare well with R1 in terms of wind anomalies and their significance. However, the distribution of moisture anomalies in CFSR is notably different. In particular, the moisture anomalies associated with the cyclonic circulation manifest a more distinct tilt, that is, a clockwise rotation of ~45°, compared to the more zonal orientation evident in R1. The CFSR moisture anomaly orientations both at 850 hPa and throughout the entire depth of the troposphere (see section 3c) are consistent with the LN08 results using Special Sensor Microwave Imager (SSM/I) total-column water vapor, suggesting that the CFSR fields are more realistic. Additionally, the CFSR data reflect a more pronounced significant shift in precipitation, with a much stronger zonal gradient of anomalous precipitation along the southeastern extent of the SPCZ. Like Fig. 2, Fig. 3 illustrates results for CFSR wind and moisture, but for the shorter period 1998–2010 so as to take advantage of available TRMM 3B42 precipitation for comparison. Despite the shorter period considered, the wind and moisture composites are quite similar. The precipitation results do not cleanly match R1 or CFSR; like R1, precipitation shows a significant southeastward expansion during the weak-inflow phase, though the spatial orientation of the precipitation is much more similar to CFSR.

<table>
<thead>
<tr>
<th>Modeling group</th>
<th>Model name</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Max Planck Institute for Meteorology (MPI)</td>
<td>Max Planck Institute Earth System Model, low resolution (MPI-ESM-LR)</td>
<td>1.88</td>
<td>1.88</td>
</tr>
<tr>
<td></td>
<td>MPI Earth System Model, medium resolution (MPI-ESM-MR)</td>
<td>1.88</td>
<td>1.88</td>
</tr>
<tr>
<td>Meteorological Research Institute</td>
<td>MPI Coupled Atmosphere–Ocean General Circulation Model, version 3 (MRI-CGCM3)</td>
<td>1.13</td>
<td>1.13</td>
</tr>
<tr>
<td>Norwegian Climate Centre</td>
<td>Norwegian Earth System Model, version 1 (intermediate resolution) (NorESM1-M)</td>
<td>1.88</td>
<td>2.50</td>
</tr>
</tbody>
</table>

* MIROC4h lacked RCP8.5 model output and was excluded from the subset despite being regridded using area averaging.
b. CMIP5 models

We also performed composite analysis on the selected subset of 17 CMIP5 models (Figs. 4 and 5). For visual clarity, wind and moisture anomalies are only plotted where they exceed 99% significance. Note that the wind vectors plotted for the MEM in Fig. 4 are significant at that grid cell in at least one direction in a simple majority (nine) of the models. The same criterion is applied to specific humidity, although this condition is more restrictive since the wind composites need only be significant in one direction. In any case, the cyclonic circulation evident in the reanalyses is robust across all models, though the precise location of the center of circulation varies (~ ±5°). The moisture anomalies are more varied in location, although the general “tilt” of the 0 g kg⁻¹ line through the center of the circulation is in good agreement with CFSR and is thus insensitive to the spread in climatological
Comparing the CMIP5 composite anomalies with the simulated precipitation climatology, we conclude that more realistic CMIP5 simulation of the January precipitation climatology does not correspond strongly to the fidelity with which the models reproduce the observed high-frequency wind and moisture composite anomalies.

The precipitation composites in Fig. 5 reveal less consistency in response than those for either wind or moisture. While the 4 mm day$^{-1}$ contours suggest many models have a similar response to R1 in terms of a simple significant southeastern expansion of precipitation during the weak-inflow phase, the shaded differences in precipitation confirm that many models do have a zonal
gradient in anomalous precipitation south of the com-
positing region, consistent with the idea of a precipitation
“shift” between phases. This complexity is perhaps most
evident in the model ensemble mean. Thus, despite the
models simulating very similar circulation and, to a
lesser extent, moisture anomalies, their precipitation
responses are highly variable, potentially related to equally
variable convection schemes and SST patterns.

c. Vertical structure

Since the models appear to capture the observed wind
and moisture anomalies, it is of interest to examine the
vertical structure of these anomalies. This is especially
important considering the potential for interactions
between upper- and lower-level influences acting in
the SPCZ region, for example, to SPCZ–synoptic

---

**FIG. 5.** Composite analysis applied to model output from the CMIP5 historical experiment. Shown are composite differences (weak inflow minus strong inflow) of precipitation (shading, mm day$^{-1}$) for a composite index of zonal wind at 850 hPa averaged over 20°–10°S, 140°–120°W. Significance for precipitation (at the 99th percentile) is denoted by dark gray shading. Included for reference are the 4 mm day$^{-1}$ precipitation contours for the weak-inflow phase (in green) and strong-inflow phase (in brown), as well as the region in which the composite index was calculated (black box).
disturbance interactions, previously explored in upper-level vorticity fields (Widlansky et al. 2011; Matthews 2012). Figure 6 highlights the vertical structure of the wind and moisture anomalies at four levels in both of the CFSR as well as the CMIP5 MEM (R1 is omitted, though it has a similar vertical structure) using the previously discussed significance metrics. The vertical structure in CFSR is remarkably consistent through 500 hPa (wind through 250 hPa), with an increase in the magnitude and area of significance of the moisture anomalies at 700 and 500 hPa compared to 850 hPa. This result is consistent with vertical moisture profiles obtained from radiosondes at Nauru, which suggest that moisture in the free troposphere increases more than the boundary layer moisture in transition from a tropical, nonconvective environment toward a convective environment (Holloway and Neelin 2009). The models are similarly consistent, with at least half agreeing on the significance of the moisture anomalies to the northeast and southwest of the cyclonic anomaly. The deep-layer equivalent barotropic structure seen here confirms that the effects of varying low-level inflow in the SPCZ are not confined to the lower atmosphere, consistent with the vertical extent of deep convection. This result is also consistent with earlier work showing a similar structure associated with the South Atlantic convergence zone (Robertson and Mechosso 2000).

**Fig. 6.** Results of the composite analysis at 850 hPa is shown at multiple vertical levels. In (d) CFSR is the same as in Fig. 2a, and MEM is the same result as in Fig. 4, though now the significance conventions of Fig. 2 are used (i.e., nonsignificant wind and moisture anomalies are still plotted). For (a) 250, (b) 500, and (c) 700 hPa, the significance is associated with each respective level [i.e., moisture in CFSR (c) is significant if it exceeds the 99th percentile of random composites generated at 700 hPa, etc.]. Also included for reference is the mean 4 mm day$^{-1}$ precipitation contour (in green), as well as the region in which the composite index was calculated (black box).
4. Lead–lag composites

a. Reanalysis

Figure 7 illustrates the development and decay of both weak-inflow and strong-inflow phases of the composite index in CFSR in terms of wind, moisture, and precipitation on three specific days; each of these days is referred to based on its timing before or after the peak of a phase; for example, two days before a peak is denoted as day $-2$, while two days after is day 2. Aside from the cyclonic anomaly associated with an eastward SPCZ shift, two other particularly broad regions of anomalous circulation are evident throughout the weak-inflow phase: an anomalous cyclone south of Australia and an anomalous anticyclone in the south-central Pacific ($-60^\circ$S, $140^\circ$W). Similar anomalies exist during the strong-inflow phase in the same regions, though they are of opposite sign and are most evident on day $-2$. Figure 8 (left panel) shows SLP anomalies from day $-5$ through day 0 for both phases. The two broad regions of anomalous circulation are confirmed in the SLP fields on day $-2$ and day 0. Of greater interest, however, is the difference in propagation of negative SLP anomalies between the two phases. During the weak-inflow phase, a negative SLP anomaly centered around $35^\circ$S, $150^\circ$W slowly propagates eastward from day $-5$ to day $-2$. From there, it drifts toward the northwest and, ultimately, becomes recognized as the cyclonic circulation south of the composite region. In contrast, a similar negative SLP anomaly exists at $40^\circ$S, $165^\circ$W on day $-5$ during the strong-inflow phase, but it drifts toward the southeast instead and effectively avoids interaction with the composite region.

The SLP composites hint at differences in the interactions of midlatitude transients with the SPCZ: we hypothesize that the Southern Hemisphere storm track [collocated with the westerly jet axis, see Nakamura and Shimpo (2004)] preceding a weak-inflow peak is oriented in such a way that it increases SPCZ–storm interaction.

FIG. 7. Lead–lag composite differences using CFSR data. Each panel shows the difference in wind (vectors, m s$^{-1}$) and specific humidity (shading, g kg$^{-1}$) between the weak-inflow/strong-inflow phase composite at a given number of days before or after a maximum or minimum (i.e., $-2$, the top row, shows composites 2 days before a maximum or minimum) and the mean 27 December–4 February (40-day period) state. Significance for wind and moisture (at the 99th percentile) is denoted by black vector color and dark gray shading, respectively. Also included for reference are the composite 4 mm day$^{-1}$ precipitation contour (weak-inflow phase, in green; strong-inflow phase, in brown), the mean 27 December–4 February (40-day period) 4 mm day$^{-1}$ precipitation contour (in light yellow), and the region in which the composite index was calculated (black box).
Figure 9 shows the composite 250-hPa wind fields for the weak-inflow and strong-inflow phases averaged between day −5 and day −3. Three regions of interest stand out in the difference between the two phases and are significant at the 99th percentile: 1) the subpolar jet (SPJ) south of Australia, 2) the reconnection of the SPJ and subtropical jet (STJ) in the central Pacific (~145°W), and 3) the westerly duct near the compositing region.
in the eastern equatorial Pacific (~100°W). The SPJ region of increased strength (~40°S, 120°E–180°), or alternatively its northward displacement, during the weak-inflow phase is collocated with the anomalous circulations south of Australia shown in Fig. 7, indicating another equivalent barotropic response. Beyond 160°E the weak-inflow phase shows two jets of similar magnitude while the strong-inflow phase has a stronger SPJ. Additionally, the STJ streak in the weak-inflow phase does not turn poleward until nearly 160°W, or ~30° farther to the east than in the strong-inflow phase. Both features hint at a northerly displaced storm track during the weak-inflow phase consistent with the SLP results, particularly if more storms follow the STJ. While the STJ is weak and not the dominant SH jet in austral summer, synoptic disturbances still propagate along it (Nakamura and Shimpo 2004). The presence of a westerly duct in the equatorial Pacific eliminates an interhemispheric barrier and allows eddies to propagate across the equator (Webster and Holton 1982; Hoskins and Ambrizzi 1993; Matthews 2012).

Additional composite analyses based on the 250-hPa zonal wind at a lead time between 3 and 5 days in each of the regions of interest outlined above were performed (not shown). Most composites show a circulation anomaly similar to that seen in Fig. 2 for CFSR, although the areal mean anomalous winds in the region bounded by 20°–10°S, 140°–120°W are consistently weaker (~20% of
the original composite value) and the characteristic anomalous circulation is broader and less organized. The moisture and precipitation responses are similarly weak. Though some of the wind and moisture anomalies are significant at the 99th percentile, few if any of these anomalies are in the vicinity of the original circulation and moisture anomalies (those seen in Figs. 2–4). As a result, direct confirmation that differences in the storm track and position of the eastern equatorial Pacific westerly duct are responsible for the original composite analysis’s hallmark circulation, moisture, and precipitation anomalies is weak.

b. CMIP5 models

Analogous to Fig. 7, Fig. 10 depicts the CMIP5 MEM (calculated by averaging day $-2$ results across all models, etc.) wind and moisture composites. There is remarkable agreement on the existence of the anomalous anticyclone in the central Pacific during the weak-inflow phase leading up to day 0 and beyond, despite the moisture anomalies showing broad disagreement. The individual models (not shown) behave similarly to Fig. 4 in that they exhibit moisture anomalies of similar magnitude to CFSR but disagree substantially on the location of these anomalies. Similarly, the strong-inflow phase shows agreement on the existence of an anomalous cyclone in the central Pacific. On day $-2$, there are hints of an anomalous cyclone (anticyclone) in the weak-inflow (strong-inflow) phases south of Australia, though they are displaced to the east compared to CFSR, and neither shows much coherence beyond day $-2$. Though the hallmark circulation and moisture anomalies associated with the weak- and strong-inflow phases (see Fig. 4) are significant at the 99th percentile in a majority of models, particularly on day 0, neither the anomalous circulations south of Australia nor those in the south-central Pacific meet the significance threshold (not shown).

Based on the connection between the anomalous circulations both south of Australia and in the south central Pacific and the strength and northward displacement of the South Pacific jet, it appears that the orientation of the feature is dissimilar enough from CFSR in many models to produce a different response or none at all. Figure 8 (right panel) confirms the central Pacific anomalies as well as the existence and eventual decay of the anomalies near Australia. Unlike CFSR, there is less of a sense that the negative SLP anomaly propagates and stalls near the SPCZ in the weak-inflow phase; instead, it appears that...
the negative SLP anomaly associated with the anomalous circulation south of the composite region simply grows in magnitude by day 0. Analysis of individual models (not shown) confirms that the MEM results are not a product of averaging; each model exhibits a similar response.

The three areas of anomalous winds outlined for CFSR at 250 hPa between the weak-inflow and strong-inflow phases related to storm tracks appear in the CMIP5 models though they are consistently weaker and displaced (not shown). Additionally, all three are rarely recognizable in an individual model. Confirming this lack of coherence, the model ensemble mean shows very weak anomalies related to the SPJ south of Australia and the westerly duct in the equatorial eastern Pacific and no anomaly in the central Pacific associated with the STJ. That the models in fact simulate varying low-level inflow further confirms that the variation of the storm track or westerly duct position is not the sole influence.

c. Vertical structure

In both CFSR and the MEM, the day 0 circulation anomaly tends to form and decay nearly simultaneously.
at all vertical levels. Specific humidity fields, shown in Fig. 11 as composite differences, do show some spatial and temporal differences. On day $-2$, both CFSR and the MEM manifest positive moisture anomalies forming in the compositing region at 500 hPa, while the response at 850 hPa is somewhat muted. By day 0, positive moisture anomalies have formed at both levels, though the anomaly at 850 hPa extends farther southeast. In contrast, the negative moisture anomaly southwest of the compositing region forms by day $-1$ and persists through day 2 at 850 hPa, though there is a more subtle response at 500 hPa. The difference in moisture response could be from differences in the anomalous source (e.g., deep convection at 500 hPa versus advection at 850 hPa), though additional work is needed to separate the components of the response.

5. Summary and conclusions

As in LN08, composite analysis of SPCZ-region wind, moisture, and precipitation data in both reanalyses and CMIP5 model output points to robust relationships between high-frequency low-level (trade wind) inflow and tropospheric moisture and precipitation fluctuations. One caveat, however, is the number of differences between R1 and CFSR; we conclude that the CFSR is a better reference for the region than R1 until an appropriate observational dataset of equal length is available at the daily time scale. The confirmation that CMIP5 models are able to reproduce the observed relationship between these variables, especially considering the diversity of climatological SPCZs across the models, suggests that biases in SPCZ simulation are not due to poor representation of this particular synoptic-scale response. However, key differences exist between the reanalysis products, specifically CFSR, and the CMIP5 models. In particular, the precipitation response is more complex in the CMIP5 models; while some models show a precipitation shift toward the east in the weak-inflow phase, others simulate an expansion of precipitation toward the southeast with little movement otherwise. In some models with overly zonal SPCZs, this expansion to the southeast gives the appearance of a more realistic SPCZ.

In terms of the vertical structure, the observed circulation and moisture anomalies in the vicinity of the compositing region manifest equivalent barotropic structure. This suggests that the influence of low-level inflow is not confined simply to the boundary layer but rather extends...
over a deep layer. In addition, the positive moisture anomalies in both CFSR and the CMIP5 model ensemble mean (MEM) above 850 hPa are stronger and, at 500 hPa in particular, may even lead the 850-hPa anomalies by 1 or 2 days. In comparison, the negative anomaly appears to be more strongly confined to the lower levels. Interpretation of these results would benefit from a better understanding of the processes that contribute to vertical moisture distribution (e.g., Holloway and Neelin 2009; Lintner et al. 2011).

The lead–lag analysis applied to CFSR as well as the CMIP5 models, which is summarized schematically in Fig. 12, suggests two prominent antecedent circulation features for weak inflow and strong inflow, namely a cyclone (anticyclone) south of Australia and an anticyclone (cyclone) in the south-central Pacific, respectively. Aloft, the strength and location of the SPJ south of Australia, the STJ streak in the central Pacific, and the westerly duct in the equatorial eastern Pacific vary with the composite index at a lead time of 3–5 days. These features together with an analysis of SLP anomalies suggest increased interaction between the SPCZ and synoptic disturbances during the weak-inflow phase. Further work is needed to understand the different mechanisms of low- and upper-level forcings at synoptic time scales, especially since the CMIP5 models differ drastically from CFSR particularly at 250 hPa in this analysis.

Another direction for future work is to gain an understanding of how these anomalies may change owing to the effects of anthropogenic warming in the region. The results of a preliminary lead–lag analysis applied to data from the CMIP5 RCP8.5 scenario are shown in Fig. 13, with the caveat that any changes shown are likely biased due to errors in precipitation and SST climatology across the South Pacific. One key difference in both weak-inflow and strong-inflow phases is the increased magnitude of the positive moisture anomalies. In addition, the zonal extent of the precipitation shift during the weak-inflow phase in the RCP8.5 MEM is approximately 5° longitude farther than that in the historical MEM (135°–115°W versus 125°–110°W) despite a projected future westward displacement of the mean January SPCZ position by 10°; in contrast, the strong-inflow phase shows even less deviation from its climatological mean than in the historical results. This result is qualitatively consistent with expectations based on warming-induced moistening of the troposphere over the South Pacific.
That is, during events of increased low-level inflow, the climatologically moister environment is less susceptible to dry air inflow shutting off deep convection; conversely, less moistening is needed during relaxed easterly events to trigger convection east of the climatological SPCZ position. Nearly all of the SLP anomalies in the RCP8.5 scenario are weakened (not shown), with the only possible exception being that associated with the weak-inflow phase anomalous cyclone south of the composite region. These weakened dynamic anomalies are consistent with a weaker Walker circulation and the resultant weaker SLP gradient across the Pacific (e.g., Vecchi and Soden 2007). An exploration of these potential changes in response using a future climate scenario with ocean SST bias adjustments akin to those discussed in Widlansky et al. (2013) would aid in verification and is planned. The changes outlined here highlight the importance of focusing on future changes beyond simple shifts in the mean state in the SPCZ region.

Though synoptic changes to low-level inflow east of the SPCZ do not appear to be a significant source of climatological SPCZ simulation biases, this does not exclude the possibility that other synoptic-scale processes, such as interactions between synoptic disturbances and eddies, are poorly represented and contribute to biases. A statistical compilation of such interactions in each model explored within this paper is planned. In addition, examining differences in the climatological position of negative zonal stretching deformation among CMIP5 models, as was done for past climate scenarios (Mantsis et al. 2013), would help to identify how accurately (in a spatial sense) these synoptic interactions occur in model simulations. The results of forcing a GCM or regional model with a correctly positioned region of negative zonal stretching deformation, as well as with a more realistic storm track, would be a significant step forward in terms of understanding the magnitude of influence these factors have on an accurately simulated SPCZ both at synoptic and seasonal time scales.

Acknowledgments. This work was funded by New Jersey Agricultural Experiment Station Hatch Grant NJ07012 and National Science Foundation Grant NSF-AGS-1312865. We acknowledge the World Climate Research Programme Working Group on Coupled Modelling, which is responsible for CMIP, and we thank the climate modeling groups (listed in Table 1) for producing and making available their model output. For CMIP the Department of Energy’s Program for Climate Model Diagnosis and Intercomparison provides coordinating support and led development of software infrastructure in partnership with the Global Organization for Earth System Science Portals. We thank Anthony Broccoli, Brian Mapes, and David Neelin for useful discussion and feedback as well as three anonymous reviewers for their comments and suggestions. We also thank Anthony DeAngelis and Paul Loik for assistance in downloading and regridding the CMIP5 model output.

REFERENCES


