Characteristics of Drought and Persistent Wet Spells over the United States in the Atmosphere–Land–Ocean Coupled Model Experiments

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ABSTRACT: Atmosphere–land–ocean coupled model simulations are examined to diagnose the ability of models to simulate drought and persistent wet spells over the United States. A total of seven models are selected for this study. They are three versions of the NCEP Climate Forecast System (CFS) coupled general circulation model (CGCM) with a T382, T126, and T62 horizontal resolution; GFDL Climate Model version 2.0 (CM2.0); GFDL CM2.1; Max Planck Institute (MPI) ECHAM5; and third climate configuration of the Met Office Unified Model (HadCM3) simulations from the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) experiments.

Over the United States, drought and persistent wet spells are more likely to occur over the western interior region, while extreme events are less likely to persist over the eastern United States and the West Coast. For meteorological drought, which is defined by precipitation ($P$) deficit, the east–west contrast is well simulated by the CFS T382 and the T126 models. The HadCM3 captures the pattern but not the magnitudes of the frequency of occurrence of persistent extreme events. For agricultural drought, which is defined by soil moisture

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(SM) deficit, the CFS T382, CFS T126, MPI ECHAM5, and HadCM3 models capture the east–west contrast.

The models that capture the west–east contrast also have a realistic $P$ climatology and seasonal cycle. ENSO is the dominant mode that modulates $P$ over the United States. A model needs to have the ENSO mode and capture the mean $P$ responses to ENSO in order to simulate realistic drought. To simulate realistic agricultural drought, the model needs to capture the persistence of SM anomalies over the western region.

**KEYWORDS:** Drought; CMIP models

1. **Introduction**

Drought is one of the major water management issues facing the western region of the United States. In recent years, the western interior region has suffered many episodes of long-term droughts. For example, the Colorado River basin suffered severe drought for 8 years from 2000 to 2007. The unregulated inflow to Lake Powell was below normal for 7 out of 8 years during that period. Under these circumstances, careful planning when knowing the future projections of water resources may mitigate the impact of drought. Most projections rely upon the simulations from atmosphere–land–ocean coupled models. For example, the Intergovernmental Panel on Climate Change (IPCC) Coupled Model Intercomparison Project phase 3 (CMIP3) experiments were performed by several coupled models with different CO$_2$ scenarios. In the Fourth Assessment Report of the IPCC, it is projected that rainfall in the United States is likely to increase over the Northeast and to decrease over the Southwest (Field et al. 2007). River streamflows will also decrease as demonstrated by the downscaling experiments over the Colorado River basin (Christensen et al. 2007). While most models project a drier Southwest, the spread among the projected streamflow is large. An important question is whether the drought and persistent wet episodes in the CMIP3 model simulations are realistic. What are the conditions that a model must capture in order to simulate realistic drought? The answers to these questions will help us to interpret model projections of water resources.

There are two hydroclimate regimes over the United States. The western interior region is very dry with an annual rainfall of less than 1.5 mm day$^{-1}$. Once drought or extreme wet events occur, they can persist for more than 9 months. Over the eastern United States, it is relatively wet and rainfall anomalies are less persistent (Mo and Schemm 2008a). Extreme events are often selected based on anomalies defined as departures from a model’s climatology. Does a model need to have a realistic climatology in order to simulate the occurrence of drought and persistent wet spells?

Drought is defined as persistent dryness. While drought can occur because of random forcing, it is often triggered and maintained by low-frequency forcing such as sea surface temperature anomalies (SSTAs) (Hoerling et al. 2009). As documented in both observational analyses (Ropelewski and Halpert 1987; Ropelewski and Halpert 1989) and model experiments by the Climate Variability and Predictability (CLIVAR) drought working group (Schubert et al. 2009), drought over the United States is triggered and maintained by El Niño–Southern Oscillation (ENSO). The Atlantic SSTAs directly influence rainfall over the Southeast, but
their largest impact is to modulate the impact of ENSO on rainfall over the United States. Mo (Mo 2011) examined the onset of the drought occurrence over the United States; she found that a cold ENSO event often occurred one season before the onset of drought over the southern Great Plains and the Gulf States. The North Pacific SSTAs are also known to impact rainfall over the western United States (Goodrich 2007). Barlow et al. (Barlow et al. 2001) identified ENSO, the North Pacific SST mode, and the Pacific decadal oscillation (PDO) as major SSTA modes that modulate summer drought. Soil moisture (SM) does not usually trigger drought, but agricultural drought is often associated with SM persistence. Does a model need to capture all SSTA modes and their associated precipitation ($P$) anomalies in order to simulate realistic drought?

The objectives of this paper are 1) to document the characteristics of drought or wet spells simulated in the atmosphere–land–ocean coupled experiments, 2) to diagnose the sources of model errors in simulating persistent extreme events and SM anomalies, and 3) to quantify the features that a model must capture in order to simulate realistic drought. Data and experiments examined are described in section 2. In section 3, the preferred areas for drought (wet spells) to occur are identified from observations and the coupled experiments. The model’s climatology is presented in section 4. The ability for the models to capture the impact of SSTA forcing on drought is examined in section 5. Conclusions and a discussion are given in section 6.

2. Data and procedures

2.1. Observations

In this study, two observed $P$ datasets are used. One is the monthly-mean $P$ dataset obtained from the University of Washington (UW), where $P$ is derived from index stations (Tang et al. 2009). The dataset covers the period from 1915 to 2007. Another $P$ dataset is the monthly-mean Climate Prediction Center (CPC) unified $P$ data (Xie et al. 2010), which are based on all station data available for a given month. This dataset covers the period from 1950 to 2010. The horizontal resolution for both datasets is 0.5°. Each dataset is treated as one member of the $P$ ensemble.

SM data are taken from the North American Land Data Assimilation Systems (NLDAS) from the UW because no long-term observed dataset is available. The SM ensemble members are outputs from three land surface models (LSMs): the Variable Infiltration Capacity model (VIC) (Liang et al. 1994; Liang et al. 1996), Noah model (Koren et al. 1999; Ek et al. 2003), and Sacramento model (SAC) (Burnash et al. 1973). The dataset covers the period from 1916 to 2006. Model descriptions and properties can be found in Wang et al. (Wang et al. 2009). All models are driven by the same forcing. The SM time series from each model simulation is considered as one member of the SM ensemble.

The SST data are the reconstructed monthly SSTs from Smith et al. (Smith et al. 1996). The version used is Extended Reconstructed Sea Surface Temperatures version 2 (ERSST v2), and the horizontal resolution is 2°. The dataset covers the period from 1910 to 2009. For all datasets, the monthly-mean climatology is the mean averaged over the study period for that month. Anomalies are defined as departures from the monthly-mean climatology.
2.2. Atmosphere–land–ocean coupled model simulations

We examine simulations from seven models. Four models are taken from the World Climate Research Programme (WCRP) twentieth-century climate change model simulations (20C3M). They are the Geophysical Fluid Dynamics Laboratory Climate Model version 2.0 (GFDL CM2.0), GFDL CM2.1 (Delworth et al. 2006), Max Planck Institute (MPI) ECHAM5, and third climate configuration of the Met Office Unified Model (HadCM3). The model outputs are available from the Program for Climate Model Diagnosis and Intercomparison (PCMDI) at their website (http://pcmdi-cmip.llnl.gov/). The detailed documentation on 20C3M experiments can also be found there. These models have realistic ENSO (Meehl et al. 2007) and were used to examine drought over the Southwest (Seager et al. 2007).

The remaining three models are from the National Centers for Environmental Prediction (NCEP) Climate Forecast System (CFS) model (Saha et al. 2006; Saha et al. 2010). The model physics and dynamics are the same, but the horizontal resolution is different. Three coupled runs were made with the horizontal resolutions of T382, T126, and T62, respectively. The model is similar to the CFS version 2 but coupled with the GFDL Modular Ocean Model, version 3 (MOM3). It is also coupled with the Noah LSM, which has four soil layers and more realistic boundary layer physics (Xia et al. 2012; Mitchell et al. 2004). For the WCRP CMIP3 experiments, most simulations run from 1800 to 2000. Some simulations have incomplete soil moisture records. While the T62 and T126 CFS runs have nearly 200 years, the T382 CFS runs are only 55 years long. Each model’s resolution, the total simulated years, and the number of ensemble members are given in Table 1.

Table 1. Coupled model experiments, total length of records and SST modes and the variance explained.

<table>
<thead>
<tr>
<th>Model</th>
<th>Resolution (lon, lat)</th>
<th>Total years (runs)</th>
<th>Trends</th>
<th>ENSO</th>
<th>ATL1</th>
<th>ATL2</th>
<th>N_PAC1</th>
<th>N_PAC2</th>
<th>ATL2-NPAC Combined</th>
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<tr>
<td>GFDL CM2.0</td>
<td>144° × 90°</td>
<td>420 (3)</td>
<td>14.2</td>
<td>6.8</td>
<td>22.0</td>
<td>4.8</td>
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<td></td>
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</tr>
<tr>
<td>GFDL CM2.1</td>
<td>144° × 90°</td>
<td>700 (3)</td>
<td>7.1</td>
<td>27.3</td>
<td>7.9</td>
<td>7.2</td>
<td></td>
<td>5.1</td>
<td></td>
</tr>
<tr>
<td>MPI ECHAM5</td>
<td>192° × 96°</td>
<td>814 (4)</td>
<td>22.4</td>
<td>7.9</td>
<td>7.2</td>
<td>5.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HadCM3</td>
<td>96° × 72°</td>
<td>280 (2)</td>
<td>14.7</td>
<td>6.7</td>
<td>3.1</td>
<td>10.3</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NCEP T382</td>
<td>1152° × 576°</td>
<td>220 (4)</td>
<td>17.7</td>
<td>7.9</td>
<td>7.2</td>
<td>5.1</td>
<td></td>
<td>11.9</td>
<td></td>
</tr>
<tr>
<td>NCEP T126</td>
<td>384° × 190°</td>
<td>193 (2)</td>
<td>17.8</td>
<td>17.3</td>
<td>6.7</td>
<td>10.0</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>NCEP T62</td>
<td>192° × 94°</td>
<td>193 (2)</td>
<td>21.8</td>
<td>17.3</td>
<td>6.7</td>
<td>15.2</td>
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<td>27.2</td>
<td>20.5</td>
<td>5.8</td>
<td>6.2</td>
<td></td>
<td>7.2</td>
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</tr>
</tbody>
</table>

2.3. Drought classification

Meteorological drought is measured by P deficit (Keyantash and Dracup 2002). The drought index used to classify meteorological drought is the 6-month standardized precipitation index (SPI6). Agricultural drought is measured by soil moisture deficit. The soil moisture anomaly percentile is used to classify agriculture drought (Mo 2008). The hydrological drought is not considered because runoff is not archived by all models.
The SPI6 is computed from two $P$ datasets, the UW for 1915–2007 and the CPC unified $P$ data for 1948–2007, by following the method outlined by McKee et al. (McKee et al. 1993; McKee et al. 1995). SM anomaly percentiles are computed for each member of the SM NLDAS outputs. The drought indices are computed for each member of the coupled model experiments following the same procedures as observations.

For a given experiment, we append the time series of SPI6 (or SM percentiles) from each member together to form one long time series. The term $N_{\text{total}}$ is the record length. It is defined as the total months of runs from a given experiment or the length of an observed dataset. At each grid point, an extreme negative (positive) event is selected when the SPI6 index or the SM percentile is below (above) a certain threshold. The threshold for SPI6 is $-0.8$ ($0.8$) for a dry (wet) event (Svoboda et al. 2002). The threshold for the SM percentile is 20% (80%). At each grid, the number of months $N$ that extreme events occur is 20% of the record length ($N/N_{\text{total}} = 20\%$). Because a drought (persistent wet event) usually means persistent dryness (wetness), a drought (wet) episode is selected when the index is below (above) this threshold for three seasons (9 months) or longer. The number of months that an extreme event persists for more than 9 months is labeled as $N_p$. The frequency of occurrence (FOC) of drought or persistent wet spells is defined as

$$\text{FOC} = N_p/N.$$  \hspace{1cm} (1)

The same procedure is used to calculate the FOC based on SM percentiles.

### 3. Persistent extreme events over the United States

The FOC for extreme positive/negative events that persisted more than 9 months based on SPI6 or SM percentiles is calculated for observations. The combined FOC for positive and negative events is given in Figure 1a for meteorological drought (SPI) and Figure 2a for agricultural drought (SM percentile). A 9-point smoother is applied to FOC before plotting. Drought and persistent wet spells are more likely to occur over the western interior states, consistent with Mo and Schemm (Mo and Schemm 2008a). For both plots, the east–west contrast is striking. For SPI6, the map is noisier because of short data records. It shows that more than 25%–35% of extreme events that persist more than three seasons are located over the interior western region and the Great Plains west of 95°W. The FOC along the West Coast and the eastern United States east of 95°W is smaller. For SM anomalies, more than 40%–50% of extreme events over the western interior states west of 95°W persist more than three seasons with maxima located over the Dakotas and Montana. Over the eastern United States, extreme events are less persistent. Notice the differences between FOC maps based on SPI6 (Figure 1a) and SM (Figure 2a) over the northwestern region, including eastern Montana, North Dakota, and Minnesota. Over the Northwest, snow is another source for SM. The accumulation of snow in winter and the timing and amounts of snowmelt in spring contribute to the SM variability. The SPI is derived from $P$ alone and it does not contain information on snowmelt.

For meteorological drought, the NCEP T382 CFS model captures the east–west contrast well. It shows that extreme events are more likely to persist over the
Figure 1. FOC of extreme events that persist more than 9 months based on SPI6 index averaged over positive and negative events for (a) observations, (b) GFDL CM2.0, (c) GFDL CM2.1, (d) MPI ECHAM5, (e) NCEP T382 CFS, (f) NCEP T126 CFS, (g) NCEP T62 CFS, and (h) HadCM3. The contour interval is 0.05. Values greater than 0.2 are shaded according to the color bar.
western interior region and the southern Great Plains (Figure 1e), but the FOC over the Southeast is too large. The FOC simulated by the T126 CFS model is similar to the T382 CFS in that it also captures the east–west contrast (Figure 1f). For the NCEP T62 CFS, the maximum of the FOC shifts southward to New Mexico and

Figure 2. As in Figure 1, but for extreme events selected according to SM.
the FOC is too weak over much of the west. The GFDL CM2.0, GFDL CM2.1, and MPI ECHAM5 models place the maximum of the FOC over the Southern Plains and the Gulf States. None of these models captures the east–west contrast. The HadCM3 does show the east–west contrast and has a maximum over the Four Corners region, but the magnitudes of the FOC are too low.

For SM, the NCEP T382 CFS, NCEP T126 CFS, MIP ECHAM5, and HadCM3 models capture the east–west contrast of the FOC and compare favorably with the NLDAS (Figure 2a). The NCEP T62 CFS also shows the east–west contrast, but values over the interior northwestern mountains are too weak. The SM anomalies for the two GFDL models do not persist enough to have persistent events.

In addition to persistence, we also examined whether the magnitudes of the extreme events simulated by models are realistic. Figure 3 shows the composite differences between positive and negative extreme events for selected models based on SPI6. Here, we only require the SPI6 to be greater than 0.8 (less than −0.8). Persistence is not a requirement. The patterns of the composite differences are similar to the P annual mean (Figure 4). Both show that the magnitudes of the extreme events are larger over the West Coast and the Southeast and smaller over the western interior region. The extreme events simulated by models have larger magnitudes than the P analysis. All models also have difficulty capturing the dryness over the western interior region. The GFDL CM2.1 shows large magnitudes over the same area where it has a maximum in FOC. Overall, the model-simulated P anomalies have larger magnitudes than observations over the western

Figure 3. Composite difference of P anomalies between positive and negative extreme events when the SPI6 is greater than 0.8 (less than −0.8) for (a) observations, (b) GFDL CM2.1, (c) NCEP T382 CFS, and (d) HadCM3. The contour interval is 0.2 mm day⁻¹.
Figure 4. Annual-mean $P$ for (a) observations, (b) GFDL CM2.0, (c) GFDL CM2.1, (d) MPI ECHAM5, (e) NCEP T382 CFS, (f) NCEP T126 CFS, (g) NCEP T62 CFS, and (h) HadCM3. The contour interval is 1 mm day$^{-1}$. Contours for 0.5 mm day$^{-1}$ are added.
interior region. The models fail to capture the FOC not because of weak anomalies but because of the lack of persistence.

The spread of both extreme events and FOCs among the models is large. Therefore, the comparison among these models gives us insight into the conditions that a model must capture in order to simulate drought or persistent wet spells. In the following sections, we will explore why some models are able to capture the observed FOC of persistent extreme events and some do not.

4. Seasonal cycle and persistence

4.1. P climatology and seasonal cycle

Even though drought and persistent wet spells are defined as persistent $P$ anomalies, a model still needs to have a realistic climatology to simulate drought. This is because the east–west contrast is a part of the seasonal $P$ climatology. The western interior region is a dry region. The vegetation is sparse and the water holding capacity is large (Wang et al. 2009). The mean rainfall is less than 1.2 mm day$^{-1}$ (Figure 4a), and the magnitude of the annual cycle, estimated by the difference between $P$ maximum ($P_{\text{max}}$) and $P$ minimum ($P_{\text{min}}$) divided by 2, is only about 0.5 mm day$^{-1}$ (Figure 5a). Both rainfall and soil moisture have a long memory and are likely to persist. Once an area is under drought, it will take a long time to recover. The eastern region is the opposite of the western interior regime. It is wet with an annual rainfall of more than 3.5–4 mm day$^{-1}$. The weak seasonal cycle implies that it rains consistently throughout the year. If drought occurs in one season, it is likely to get relieved a few seasons later, making extreme events less likely to persist (Mo and Schemm 2008b).

The models that capture the east–west contrast of the FOC also have a realistic seasonal cycle and climatology. The annual-mean $P$ and the seasonal cycle simulated by the models are given in Figures 4b–h and 5b–h, respectively. The NCEP T382 CFS has a realistic seasonal cycle and climatology. It also has the best performance in simulating the FOC of extreme events. Overall, the T126 CFS captures the dryness and weak seasonal cycle over the western interior region west of 105°W, but it is slightly wetter than the T382 CFS over the western mountains. Both models overestimate the annual rainfall and simulate a stronger seasonal cycle over the Southeast than $P$ analysis. Because the Southeast is a wet region, the excessive rainfall does not have large impact on the FOC.

The GFDL CM2.0, GFDL CM2.1, and NCEP T62 CFS models are too wet and the seasonal cycle is too strong over the western interior region. Their annual rainfall is more than 1.5–2 mm day$^{-1}$, and the seasonal cycle is about 1.5 mm day$^{-1}$. It rains often. If an area is under drought, rainfall over the next few seasons may relieve the situation. The MPI ECHAM5 and HadCM3 models are drier and have a weaker seasonal cycle over the western interior region. The driest areas over the Southwest are also the areas that have the largest FOC values.

For a model to capture the east–west contrast of the FOC, the $P$ climatology needs to distinguish dryness over the western interior region and wetness over the eastern United States. Both regimes have a weak seasonal cycle. If the seasonal cycle is too strong over the western interior region, then drought is less likely to persist because of the possibility of increased rainfall during the following season.
Figure 5. As in Figure 4, but for $P$ seasonal cycle defined as the climatological difference $(P_{\text{max}} - P_{\text{min}})/2.$
4.2. SM persistence

Even though soil moisture does not trigger drought, the persistence of soil moisture will prolong agricultural drought. We use the characteristic time $T_o$ to measure this persistence (Trenberth 1984; Mo and Schemm 2008a). It is computed from the autocorrelation $R(i)$ at lag $i$ month for $i = 1–30$,

$$T_o = 1 + 2 \sum_{i=1}^{N} (1 - i/N)R(i),$$

where $N = 30$.

Here, $T_o$ is computed for each NLDAS member. The ensemble-mean $T_o$ averaged over all members is given in Figure 6a. It shows the east–west contrast. Here, $T_o$ is less than 6 months east of 95°N and along the West Coast, but it is longer than 24 months (2 years) over the western interior region. This suggests that SM over the western interior region is more persistent than over the eastern United States. The east–west contrast also appears in the $T_o$ calculated for SPI6 (not shown). There is a consistency between $P$ and SM anomalies.

There is a good one-to-one correspondence between $T_o$ (Figure 6) and the FOC of SM (Figure 2) for observations and model simulations. Large FOC is located over the areas where $T_o$ is longer than 16–20 months. SM anomalies from the GFDL CM2.0 and CM2.1 do not persist. For these models, $T_o$ is less than 6 months over the western region and the maxima are about 8 months over the central eastern United States and the Southwest. This is consistent with weak FOC for the GFDL models. The MPI ECHAM5 captures the east–west contrast of $T_o$ but underestimates the values of $T_o$ in comparison to NLDAS. The HadCM3 has large $T_o$ over the Southwest where the SM has large FOC.

All NCEP models show large $T_o$ over the western region. The $T_o$ from the T382 CFS compares well with the ensemble NLDAS. The $T_o$ is larger than 24 months over the western interior region and less than 6 months over the east. The T126 CFS also shows the east–west contrast, but it slightly underestimates $T_o$ over the mountains. The $T_o$ from the T62 CFS is similar to that from the HadCM3. The T62 model captures the east–west contrast, but it underestimates the $T_o$ over the western interior mountains.

The close correspondence between $T_o$ and the FOC of SM suggests that agriculture drought is related to the persistence of SM. Therefore, a model needs to have a realistic LSM in order to capture agricultural drought.

5. SST forcing and impact simulated by models

5.1. SST modes

Although a model needs to have a realistic $P$ climatology and seasonal cycle to simulate realistic drought, the seasonal cycle does not trigger drought. It merely provides favorable conditions for $P$ anomalies to persist. Over the United States, drought is modulated by ENSO and the Atlantic and North Pacific SSTAs (Schubert et al. 2009; Mo et al. 2009). In this section, we identify the SSTAs modes that a model needs to have in order to simulate realistic drought characteristics. In
Figure 6. Characteristic time $T_o$ for SM for (a) NLDAS, (b) GFDL CM2.0, (c) GFDL CM2.1, (d) MPI ECHAM5, (e) NCEP T382 CFS, (f) NCEP T126 CFS, (g) NCEP T62 CFS, and (h) HadCM3. The contour interval is given by the color bar. The unit is 1 month.
addition to having these modes in their simulated SSTAs, a model also needs to capture the $P$ responses to these modes.

SSTA modes are identified by performing rotated EOF (REOF) analyses on annual-mean SSTAs. For each model experiment, we pool the annual-mean SSTAs from all members together to form one continuous time series. A REOF analysis is then performed on the model’s annual-mean SSTAs. The procedure is the same as the one described by Schubert et al. (Schubert et al. 2009). The modes and percentages of variance explained are listed in Table 1.

For observations, the first two REOFs for annual-mean SSTAs are similar to the modes found in the CLIVAR drought working group experiments (Schubert et al. 2009). The first REOF explains 27.2% of the variance (Figure 7a). It is the trend mode with positive loadings over the three southern oceans. Drought and wet spells over the western mountains and Southern Plains are often influenced by long-term trends (Mo and Schemm 2008a; Groisman et al. 2004). For these regions, the trend mode enhances the persistence of $P$.

The major forcing to modulate drought over the United States is ENSO, which is the second REOF mode and explains 20.5% of the variance (Figure 7b). It shows positive loadings over the central Pacific and along the U.S. West Coast. There are also weak negative loadings in the North and South Pacific. The pattern also has a projection onto the PDO mode. The time series associated with this mode has both interannual and decadal components. When the SSTAs are filtered to retain the frequencies less than 6 years, the first mode is the ENSO mode in the interannual frequency band (Figure 8a). Positive loadings are confined to the tropical band along the equator.

For the remaining REOFs, the order of occurrence, the percentage of variance explained, and the detailed features of the REOF depend on the datasets and the time period used to compute the REOFs. Two SST modes in the Atlantic (Figures 7c,d) are orthogonal with each other. They represent the evolution of SSTAs in the Atlantic (Deser and Blackmon 1993). The Atlantic SST mode 1 is the mode used by the CLIVAR experiments (Schubert el al. 2009). On the decadal time scales, it is known as the Atlantic multidecadal mode [Atlantic multidecadal oscillation (AMO)]. It modulates the ENSO impact on $P$ over all of the United States (Schubert el al. 2009). The two SST modes in the Pacific (Figures 7e,f) have large loadings in the North Pacific. The first North Pacific mode (Figure 7e) resembles the North Pacific mode (Barlow et al. 2001) and the second mode has a large projection onto the PDO (Figure 7f). The PDO modulates the impact of ENSO on $P$ over the western region in winter (Goodrich 2007). Both modes have impact on droughts over North America in summer (Barlow et al. 2001).

The SSTA modes and the variances explained for each model simulation are given in Table 1. All modes have the ENSO mode; however, the percentage of explained variances and details of the structure vary from one model to another. The SSTAs from model simulations are not filtered. However, the simulated ENSO modes are similar to the observed ENSO mode in the interannual band (Figure 8a). The positive loadings are confined to the tropical band ($10^\circ$S–$10^\circ$N). They differ in the westward extent. The SSTAs in the GFDL models, HadCM3, and MPI ECHAM5 extend to the Maritime Continent as noted by Joseph and Nigam (Joseph and Nigam 2006). The SSTAs in all NCEP CFS models extend only to the dateline as seen in observations. Joseph and Nigam (Joseph and Nigam 2006) also noted that the WCRP CMIP3 models have difficulty in modeling the evolution of ENSO.
For the NCEP T382 CFS model, the ENSO mode dominates which explains about 33% of the total variance. This is caused by one of the runs having a very regular oscillation between warm and cold ENSO with a 4-yr cycle. In addition to ENSO, it has a mode similar to the trend mode with positive loadings over three southern oceans. This mode explains 19.2% of the total variance. The third mode is the combination of the Atlantic mode 2 (Figure 7d) and the North Pacific mode 1 (Figure 7e). It has negative loadings over both the North Atlantic and the North Pacific. This combined mode appears in all NCEP models. The CFS models are not able to isolate multidecadal variations like the AMO or PDO.

The SST modes for the NCEP T126 CFS are similar to the T382 CFS. The first mode for the T62 CFS is the trend mode, the second mode is the ENSO mode, and the third mode is the combined mode with negative loadings over the North Pacific and the Atlantic.

The ENSO mode is the first mode for all WCRP CMIP3 models, which explains 14%–27% of the total variance (Figure 8). The MPI ECHAM5 and HadCM3 models, which have longer simulation length, both have an Atlantic mode similar to the AMO mode with a decadal cycle. They also have a mode similar to the PDO,

Figure 7. Rotated EOFs for (a) trend mode, (b) ENSO mode, (c) Atlantic SST mode 1 (AMO), (d) Atlantic SST mode 2, (e) North Pacific mode 1, and (f) North Pacific mode 2. The contour interval is 4 nondimensional units. Contours of –2 are added.
but it does not have clearly defined decadal cycles. Because ENSO is the dominant mode and has strong influence on drought over the United States, models that do not have ENSO are unlikely to capture realistic drought or persistent wet spells over the United States. In the next section, we will examine the ability for the
model to simulate the atmospheric responses to ENSO and to capture the associated precipitation.

5.2. Responses to the SST forcing

The $P$ responses to ENSO are investigated using composites. For each model, the Niño-3.4 index is computed by averaging SSTAs over the region (5°S–5°N, 170°–240°E). The composites of $P$ and SPI6 are prepared for four seasons separately based on the normalized seasonal mean of SSTAs for the Niño-3.4 index. For observations, Figure 9a shows the mean composite SPI6 difference between cold and warm ENSO events averaged over four seasons with dryness (wetness) over the Southwest and the Great Plains during cold (warm) ENSO events. These are also areas where extreme events are most likely to persist (Figure 1a). This suggests the importance of a model being able to capture the ENSO impact on SPI6 in order to simulate drought over the United States. Figure 9a also shows weak anomalies over the Gulf Coast and Florida but few statistically significant anomalies over the eastern United States. There are two reasons that the $P$ anomalies do not persist over the Southeast and the East Coast. First, because of the weak $P$ seasonal cycle and wetness over the region (Figure 5a), the lack of rainfall in winter can be compensated by summer rainfall or vice versa. Second, ENSO has an opposite impact on rainfall for summer and winter over the East Coast and the Southeast (Mo and Schemm 2008b). For winter months January–March (JFM), the composite difference between cold and warm events (Figure 10a) shows dryness over California, the Southwest, the Great Plains and the East Coast and wetness over the Pacific Northwest and the Ohio Valley. During the summer months July–September (JAS), the composite difference (Figure 10c) indicates wetness over the East Coast. Some of this rainfall can be attributed to tropical storms that a coarse resolution model is unlikely to capture.

There is a close association between model errors in capturing the impact of ENSO on SPI6 and the ability for a model to capture the FOCs of extreme $P$ events. To demonstrate this, the ENSO composites of SPI6 are prepared the same way as observations for each model (Figure 9). The areas where the composite is statistically significant at the 5% level are shaded. For the CM2.0 GFDL and MPI ECHAM5 models, the composites show negative SPI6 values located over the Gulf States and the Southeast. These are also areas with large FOC values (Figures 1b–d). Both the NCEP T382 and T126 CFS models compare well with the observations, but the magnitudes of the composite for the T382 CFS are much too weak. The NCEP T62 CFS shows large negative SPI6 values over the Southwest, but positive values over the Northwest are too strong. Also, the negative values over the Southwest do not stretch far enough north. The HadCM3 captures the pattern of the observed SPI6 composite well, except it does not show significant anomalies at latitudes north of 38°N. Unfortunately, ENSO only explains 15% of the total variance of the HadCM3. Therefore, although it is able to simulate a realistic pattern of the FOC, the magnitudes are too weak. For the GFDL models, the composites of SPI6 are too weak and are confined to the Gulf States and the Great Plains. The impact does not extend far enough to the Southwest. The similarity between the pattern of the SPI6 ENSO composite and the FOC suggests that ENSO plays a major role in modulating persistent events. A model must exhibit ENSO
Figure 9. Composite of SPI6 difference between cold and warm ENSO events averaged over four seasons for (a) observations, (b) GFDL CM2.0, (c) GFDL CM2.1, (d) MPI ECHAM5, (e) NCEP T382 CFS, (f) NCEP T126 CFS, (g) NCEP T62 CFS, and (h) HadCM3. Contour interval is given by the color bar. Areas where anomalies are statistically significant at the 5% level are shaded.
Figure 10. Composite of $P$ anomalies between cold and warm ENSO events from observations for the (a) JFM, (b) AMJ, (c) JAS, and (d) OND seasons. The contour interval is given by the color bar. The unit is 1 mm day$^{-1}$. (e)–(h) As in (a)–(d), but for the NCEP T382 CFS. Areas where anomalies are statistically significant at the 5% level are shaded.
and be able to capture the impact of ENSO on SPI6 in order to simulate realistic
drought and persistent wet spells.

It is interesting that the ENSO composite for the T382 CFS has a realistic
pattern, but magnitudes of anomalies are too weak. When broken into four seasons,
the composites of $P$ have errors (Figures 10e–h). For example, the JFM composite
does not capture positive anomalies over the Pacific Northwest, and the negative
anomalies over the Great Plains eastward are too weak. For October–December
(OND), the composite correctly captures negative anomalies over the Great Plains
and the Southeast, but the magnitudes are again too weak. The weak negative
anomalies over the Great Plains for JFM and OND are compensated by overly
strong negative anomalies in April–June (AMJ). This indicates that, in order to
simulate realistic FOC for drought or persistent wet spells, the model only needs to
capture the mean responses of $P$ during ENSO.

All models have an ENSO signal in the tropics, but not all models capture the $P$
response to ENSO. This suggests that to have the ENSO mode alone is not enough
for a model to capture the FOC for drought and persistent wet spells. The model
also needs to capture the tropical rainfall and the atmospheric responses to ENSO.
This is illustrated by showing composite differences between cold and warm
ENSO events of rainfall and 200-hPa streamfunction anomalies with the zonal
means removed for selected models (Figures 11, 12). These models are the NCEP
T382 CFS, HadCM3, and GFDL CM2.1 models. The NCEP T382 CFS has the
highest resolution and it captures the realistic SPI responses, but magnitudes of
composites are weak (Figure 9e). The HadCM3 has the most realistic SPI re-
sponses to ENSO (Figure 9h), and the GFDL CM2.1 does not capture the SPI
response realistically.

For observations, the 200-hPa streamfunction anomalies are taken from the
NCEP–National Center for Atmospheric Research (NCAR) reanalysis from 1950
to 2009. The $P$ anomalies are from CMAP from 1979 to 2009. There is no global $P$
dataset before 1979. The $P$ composite in the tropics shows a typical ENSO pattern
with suppressed $P$ over the tropical central Pacific and enhanced precipitation over
the western Pacific for winter (Figure 11a). Negative $P$ anomalies located south and
north of the suppressed convection are due to the Hadley circulation (Rasmusson
and Mo 1993). Positive anomalies located over northeastern Brazil are due to the
Walker circulation. The composite of the 200-hPa streamfunction anomalies show
a couplet straddling the equator at the location of suppressed $P$ anomalies centered
at 150°W with a wave train to North America. The wave train is the tropical
Northern Hemisphere teleconnection pattern with positive anomalies over the
North Pacific, negative anomalies over Canada, and positive anomalies over the
Southeast (Figure 11a). For summer, the convection pattern is similar to winter, but
the positive anomalies are located over Central America (Figure 12a). The com-
posite of the streamfunction anomalies shows a wavenumber-1 pattern. There is a
couplet straddling the equator in the central Pacific and another couplet with the
opposite phase in the Atlantic.

All models show the suppressed convection over the tropical central Pacific and
enhanced convection over the western Pacific and northern Brazil, but the mag-
nitudes and the detailed structures differ (Figure 11). The common errors of model
simulations are that the suppressed convection in the central Pacific extends too
far west, and the positive anomalies north and south of the suppressed rainfall
Figure 11. Composites of $P$ and 200-hPa eddy streamfunction anomalies between cold and warm ENSO events for JFM from (a) observations, (b) NCEP T382 CFS, (c) GFDL CM2.1, and (d) HadCM3. The composite for $P$ is colored and the contour interval is given by the color bar. The streamfunction is contoured every $10 \times 10^6$ m$^2$ s$^{-1}$. 
Figure 12. As in Figure 11, but for JAS.
anomalies in the central Pacific are too strong. The westward extension of the suppressed convection appears to be common in the CMIP3 models (Joseph and Nigam 2006), which is consistent with the westward extension of the SSTAs. All models capture the 200-hPa streamfunction anomaly couplet straddling the equator and a wave train downstream to North America. The phase and the magnitudes of the anomalies over the Pacific North American region differ, and they determine the responses of the precipitation anomalies.

For the NCEP T382 CFS, the positive anomalies centered at 40°N along the Atlantic coast are too strong, and the negative anomalies are located too far north. This causes weaker rainfall anomalies over the Southern Plains in comparison to observations (Figure 10e). The GFDL model responses, as indicated by Joseph and Nigam (Joseph and Nigam 2006) and Wittenberg et al. (Wittenberg et al. 2006), are too strong over the Pacific–North American region, and the positive anomalies over the Pacific Northwest extend too far inland. The most realistic responses are simulated by the HadCM3. It captures the location of the couplet and the wave train well. The HadCM3 also has the most realistic SPI responses to ENSO (Figure 9h).

The simulations for JAS are less realistic (Figure 12). All models show the westward extension of suppressed convection in the tropics, along with the location of the couplet. For the NCEP T382 CFS and the HadCM3, the center of the couplet shifts to the date line. For the GFDL CM2.1, the couplet shifts almost 60° to the west. For the NCEP T382 CFS, the enhanced convection over Central America is located too far south.

Overall, the winter simulations are more realistic than summer. Because ENSO events usually last for more than one season, the errors in summer and other seasons will limit the ability of the models to capture the low-frequency responses to ENSO. The model needs to have realistic convection over the tropics and atmospheric responses to rainfall for all seasons in order to capture the \( P \) responses to ENSO over the United States. The HadCM3 performs the best. It captures the responses to ENSO realistically. However, the ENSO mode only explains less than 15% of variance. Therefore, the HadCM3 captures the pattern of the FOC, but the magnitudes are weak. The T382 CFS has errors, but it captures the mean rainfall anomalies associated with ENSO, although the magnitudes of anomalies are weak. The high-resolution T382 CFS has the possibility to resolve terrain-related features and capture regional responses.

6. Conclusions and discussions

In this paper, we examine the extreme persistent events of rainfall simulated by seven atmosphere–land–ocean coupled models. Four models are taken from the twentieth-century climate change model simulations. Two models are from GFDL: CM2.0 and CM2.1. The other two models are the MPI ECHAM5 and the HadCM3. The remaining three models are from NCEP. These models have the same physics and dynamical structure, but they have different horizontal resolutions: T382, T126, and T62.

There are two hydroclimate regimes over the United States. The western interior states are drier and have larger water holding capacity. Therefore, drought and persistent wet spells are more likely to occur. The eastern United States is a wet regime where extreme events usually are less persistent. The coupled model
simulations need to distinguish these two regimes in order to simulate realistic
drought or persistent anomalies over the United States. For meteorological
drought, the east–west contrast is well simulated by the T382 and T126 CFS
models. The HadCM3 captures the pattern but values of FOC are too weak. For
agricultural drought, the T382 CFS, T126 CFS, MPI ECHAM5, and HadCm3
models all capture the east–west contrast.

The models that capture the east–west contrast also have a realistic climatology
and $P$ seasonal cycle. These models show a realistic $P$ climatology with a dry
western interior region and wet eastern region. The seasonal cycle for both the
western interior states and the eastern United States are weak.

Because drought or persistent wet spells are forced by SSTAs and ENSO is the
dominant mode, the model needs to have realistic ENSO mode and also success-
fully simulate the circulation and SPI6 responses to ENSO in order to capture
realistic drought. For agricultural drought, the model needs to simulate the per-
sistence of SM anomalies. This suggests the importance of a good LSM model for
coupled simulations. The decadal oscillations such as the AMO and PDO modulate
the impact of ENSO on rainfall, but the direct influence is not as strong as ENSO.
Therefore, models unable to isolate these modes can still have a chance to simulate
realistic drought.

The three NCEP models have the same physics but different resolution. Overall,
both the T126 CFS and T382 CFS are able to capture the east–west contrast and
simulate realistic FOC for persistent dry and wet anomalies. The T62 CFS captures
the east–west contrast, but the FOC values are too weak and the maximum is
shifted southward. This suggests that drought is influenced by local forcing in
addition to the global SSTAs and $P$ climatology. A coarse resolution may not be
able to capture the physics associated with the slope of mountains and regional
forcing in order to simulate realistic drought.

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