The summer Asia–North America teleconnection and its modulation by ENSO in Community Atmosphere Model, version 5 (CAM5)

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Abstract
Seasonal forecasts of summer continental United States (CONUS) rainfall have relatively low skill, partly due to a lack of consensus about its sources of predictability. The East Asian monsoon (EAM) can excite a cross-Pacific Rossby wave train, also known as the Asia–North America (ANA) teleconnection. In this study, we analyze the ANA teleconnection in observations and model simulations from the Community Atmospheric Model, version 5 (CAM5), comparing experiments with prescribed climatological SSTs and prescribed observed SSTs. Observations indicate a statistically significant relationship between a strong EAM and increased probability of positive precipitation anomalies over the US west coast and the Plains-Midwest. The ANA teleconnection and CONUS rainfall patterns are improved in the CAM5 experiment with prescribed observed SSTs, suggesting that SST variability is necessary to simulate this teleconnection over CONUS. We find distinct ANA patterns between ENSO phases, with the La Niña-related patterns in CAM5 disagreeing with observations. Using linear steady-state quasi-geostrophic theory, we conclude that incorrect EAM forcing location greatly contributed to CAM5 biases, and jet stream disparities explained the ENSO-related biases. Finally, we compared EAM forcing experiments with different mean states using a simple dry nonlinear atmospheric general circulation model. Overall, the ANA pattern over CONUS and its modulation by ENSO forcing are well described by dry dynamics on seasonal-to-interannual timescales, including the constructive (destructive) interference between El Niño (La Niña) modulation and the ANA patterns over CONUS.

1 Introduction
The agricultural sector, water resource managers, and other preparedness agencies desire monthly-to-seasonal rainfall forecasts for their decision-making, including for many vulnerable regions of the continental United States (CONUS), such as the western US and Great Plains. However, there has been marginal success in providing reliable long-range forecasts of precipitation in the warm season (Becker et al. 2014; Slater et al. 2016; Hao et al. 2018; Malloy and Kirtman 2020), mostly due to relatively weak atmospheric flow as well as weaker signals from El Niño–Southern Oscillation (ENSO) and the Madden–Julian Oscillation (MJO; Trenberth et al. 1998; Zhou et al. 2012; Tian et al. 2017; Jha et al. 2019; Hu et al. 2020). For these reasons, there is amplified interest within the scientific community to understand causes of CONUS summer hydroclimate variability and its predictability.

Recent studies have indicated that quasi-stationary Rossby waves might be a key to summertime predictability, especially for climate extremes (Liu et al. 1998; Ciancarelli et al. 2014; Lopez et al. 2019; DeAngelis et al. 2020; Lee et al. 2011; Schubert et al. 2011; Zhao et al. 2018; Beverley et al. 2019; Mariotti et al. 2021; Jong et al. 2021). Rossby waves can be excited by remote sources, often guided by westerly jet streams as equivalent barotropic structures (Lau and Weng 2002; Lee et al. 2009; Schubert et al. 2011; Moon et al. 2013; Zhu and Li 2016; O’Reilly et al. 2018; Li et al. 2021). These circumglobal patterns are a prominent mode of upper-level circulation variability (Ding and Wang 2005; Ding et al. 2011; Weaver et al. 2016); in fact, Lee et al. (2011) estimated that up to 35% of the interannual variability in the mid-latitudes can be attributed to near-circumglobal teleconnections. The causes of quasi-stationary Rossby wave activity include distinct and joint influences from the Asian summer monsoon (ASM) and ENSO, among other possible forcing.
The East Asian summer monsoon (hereby EAM and assuming summer component) is considered the northeastern branch of the ASM that affects eastern China, Korea, and Japan (Ha et al. 2018). EAM precipitation and circulation is typically out-of-phase with the western North Pacific monsoon (WNPM; Moon et al. 2013; Zhao et al. 2015), another sub-system of the ASM located south of the EAM that affects southeast Asia and the Philippines. Together, the EAM and WNPM comprise a northward-propagating rain belt with variability on intraseasonal, seasonal, interannual, and interdecadal timescales (Wang et al. 2001, 2008; Moon et al. 2013; Zhao et al. 2015; Lee and Wang 2016).

Because of its proximity to the East Asian jet, EAM heating is favorable for generating Rossby wave responses (Lau and Weng 2002; Moon et al. 2013; Zhu and Li 2016, 2018; Zhao et al. 2018). This upper-level wave pattern, the Asia–North America (ANA) teleconnection, often traverses the North Pacific and reaches North America (Lau and Weng 2002; Zhu and Li 2016; Yang et al. 2020). Zhu and Li (2016) found that ANA is associated with a northwest-southeast US dipole of positive–negative rainfall anomalies. A series of AGCM experiments in their study suggested that the EAM can force the ANA independent of ENSO. There is a robust statistical and dynamical relationship between East Asian and North American climate via this barotropic Rossby wave train (Wang et al. 2001; Moon et al. 2013; Zhao et al. 2018; Lopez et al. 2019; Yang et al. 2020), though influence is most evident in the Great Plains and Midwest regions (Lopez et al. 2019). The EAM is considered a regional manifestation of the circumglobal teleconnection (CGT), a zonal wavenumber-5 pattern in the Northern Hemisphere that is primarily influenced by Indian summer monsoon heating (Lau et al. 2004; Ding and Wang 2005; Ding et al. 2011). The CGT is associated with significant summertime temperature and rainfall anomalies across the globe, including over North America (Lau et al. 2004; Beverley et al. 2019, 2021).

The link between ENSO-forced North Pacific wave patterns and summer CONUS hydroclimate is more complex. There is evidence that the warm phase of ENSO can excite the Pacific-North America (PNA) teleconnection in the summer by itself (Schubert et al. 2002; Pegion and Kumar 2010; Hu and Feng 2012; Zhu et al. 2013; Weaver et al. 2016), but results are inconsistent. For example, Ciancarelli et al. (2014) showed that the influence of tropical Pacific SST forcing is prominent in the early summer months, whereas Weaver et al. (2009) suggested influence peaks in the late summer months.

Rather, the relationship between ENSO and CONUS hydroclimate may be more indirect; ENSO affects EAM variability and/or the mean state in which EAM forcing responses propagate. On the seasonal-to-interannual timescale, studies indicate that transitioning or developing ENSO phases impact East Asian jet variability (Ting and Wang 1997; Du et al. 2016) and/or EAM and WNPM precipitation variability (Wang et al. 2001, 2008; Wu et al. 2009; Zhao et al. 2015; Jong et al. 2021). Ding et al. (2011) found that ENSO can modulate the strength of the EAM, altering North Pacific wave patterns. On the intraseasonal timescale, the phase of ENSO can impact the period of the northward-propagating mode (Liu et al. 2016). Finally, long-range predictability from EAM-related responses may come from ENSO (Lee et al. 2011; Liu et al. 2019; Zhou et al. 2020). Potential predictability of the EAM and its teleconnections have been linked to interdecadal ENSO variability, i.e. there are decades when ENSO has a greater influence on the EAM (Song and Zhou 2015; Zhu and Li 2018; Li et al. 2019).

There are several studies on the assessment of summer teleconnections and associated North American hydroclimate in current global climate models, many of which demonstrate that SST variability is vital for representing atmospheric circulation responses over CONUS (Weaver et al. 2009; Hu and Feng 2012; Burgman and Jang 2015). Sheffield et al. (2013) found that models in Coupled Model Intercomparison Project, version 5 (CMIP5) could not simulate CONUS summer teleconnections and mechanisms related to its forced variability from SST anomalies well. More specifically, North Pacific Rossby wave patterns related to a transitioning Niño-to-Niña and western Pacific heating were weaker and inconsistent in Community Atmosphere Model, version 5 (CAM5) without air-sea interaction (Jong et al. 2021). Air-sea interaction improves EAM precipitation-SST connection (Liu et al. 2019; Islam et al. 2013; Islam and Tang 2017), though there is realistic depiction of monsoonal precipitation climatology and interannual variability in both CAM5 (Islam et al. 2013; Islam and Tang 2017) and CMIP5 (Sperber et al. 2013) overall.

Despite the perceived importance of Pacific SST variability, idealized atmospheric model experiments with a prescribed summertime climatological background state from Zhu and Li (2016) and Lopez et al. (2019) also produced the observed ANA pattern. In addition, there are few studies that explicitly assess the ANA teleconnection in global climate models, including the evaluation of predictability from prescribed SST variability. Therefore, our paper will address the lack of understanding about the ANA teleconnection in observations as well as in a state-of-the-art climate model. Using a simple framework, we isolated the EAM-forced and ENSO-modulated circulation responses and precipitation impacts over CONUS. Due to emergent literature on the important role of the mean or background state in climate models and skill in teleconnection representation and prediction (Henderson et al. 2017; Kim et al. 2020; Wang et al. 2020), we diagnosed the roles of mean state and forcing in producing the ANA patterns and related model biases on the long-range timescale.
Section 2 contains the description of the datasets, model experiments, and analysis methods for the study. Results begin in Sect. 3, with the comparison of the ANA pattern between observations and CAM5, including how ENSO influences this pattern. In Sect. 4, we explain the circulation between observations and CAM5, including how ENSO given in Sect. 5.

2 Data and methods

2.1 Observational datasets

All observations are monthly values in the June–July–August (JJA) season for 1979–2019. Zonal wind, meridional wind, geopotential height, temperature, and divergence data were obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) fifth-generation reanalysis (ERA5), which is provided on a 0.25° latitude/longitude grid (Hersbach et al. 2020). Precipitation data were taken from Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP), which combines satellite and gauge data on a global 2.5° latitude/longitude grid (Xie and Arkin 1997). Global SST data were taken from the Extended Reconstructed Sea Surface Temperature, version 5 (ERSST-5), which is provided on a 2° latitude/longitude grid (Huang et al. 2017).

2.2 CAM5 experiments

CAM5 is a global circulation model (GCM) and the atmospheric component of the Community Earth System Model (CESM), developed by the National Center for Atmospheric Research (NCAR; Neale et al. 2010). Atmosphere-only 30-year simulations were conducted with prescribed global SSTs on an ocean data model. We analyzed monthly JJA output data with a resolution of 0.47° latitude by 0.63° longitude and 30 vertical levels.

To investigate the role of SST variability on the ANA teleconnection, we performed two experiments: one simulation with prescribed climatological SSTs (hereby referred to as CAM5_climoSST), and the other simulation with prescribed observed SSTs (CAM5_obsSST). The CAM5_obsSST experiment used the 1972–2001 period for SST data. This introduces global SST variability (e.g. ENSO), compared to the CAM5_climoSST experiment, where the prescribed time-mean SST data filters out this variability.

2.3 Linear steady-state quasi-geostrophic solutions

We explored the steady-state quasi-geostrophic (QG) linearity of the response to prescribed divergence and mean state. This method starts with the linear QG vorticity equation:

\[
\frac{\partial \zeta_g}{\partial t} + v_g \cdot \nabla (\zeta_g + f) = -f_0 \nabla \cdot v
\]

(1)

where \(\zeta_g\) is the geostrophic vorticity, \(v_g\) is the geostrophic velocity, \(f\) is the Coriolis parameter, \(f_0\) is the Coriolis constant, and \(\nabla \cdot v\) is the divergence. Focusing on the steady-state response, using the mid-latitude beta-plane approximation (\(f = f_0 + \beta v\)), and viewing the divergence term on the right-hand side as forcing (denoted as \(F\)), the equation becomes:

\[
u_g \frac{\partial \zeta_g}{\partial x} + \beta v_g = F
\]

(2)

The first term on the left-hand side is the advection of geostrophic relative vorticity by the geostrophic zonal wind, and the second term on the left-hand side is the advection of planetary vorticity by the geostrophic meridional wind. These terms “compete” to balance the divergence forcing term on the right-hand side. Finally, we linearize the equation about a prescribed time mean geostrophic zonal wind and solve for the geostrophic meridional wind via a fixed-point iteration. The calculated meridional wind field is converted to a geopotential height field.

In using this equation, we are interested in how differences in the mean state, \(\overline{U}\) (hereby referred to as \(U\)) and EAM-related divergence forcing, \(F\), alter the linear QG response, which is the ANA pattern in this study. We specified \(\overline{U}\) as the 250-hPa zonal wind (U250) climatology and the divergence forcing (\(F\)) as the strong–weak EAM 250-hPa divergence (DIV250) anomaly composite. These values are taken from observations, CAM5_climoSST, or CAM5_obsSST. We also considered \(\overline{U}\) and \(F\) that are conditioned on ENSO state (Niño or Niña) to assess the importance of indirect ENSO modulation of responses and biases. Finally, we tested direct ENSO modulation by adding tropical Pacific El Niño or La Niña divergence forcing and subtracting that response from response with only EAM forcing.

2.4 Dry AGCM experiments

To isolate the dry atmospheric processes, we also tested EAM forcing with a dry, baroclinic, nonlinear AGCM. This is a spectral model with Rhomboidal truncation at R42, which equates to a 1.7° latitude by 2.8° longitude grid, and it has 26 vertical levels. The model solves the full primitive
equations of divergence, vorticity, temperature and surface pressure. The AGCM is adapted from Brenner et al. (1984) to remove moist processes. Newtonian cooling is specified throughout the troposphere with enhanced damping near the surface. Rayleigh friction is specified at the lower levels and mimic realistic land-sea frictional contrasts. Versions of this dry AGCM have been used in Kirtman et al. (2001) and is described in detail in Brenner et al. (1984). This simple, idealized model is used for evaluating the large-scale teleconnections, primarily quasi-stationary wave activity and dry dynamic processes.

The model requires a mean state as input, with an option to add heating forcing. The JJA surface temperature climatology is input as mean state for the model. Climatology was calculated from ERA5, CAM5_climoSST, and CAM5_obsSST data and interpolated to the model’s grid. Each experiment with the respective mean state was integrated forward for 900 days, but analysis excluded the spin-up period of 100 days. EAM-forced responses are calculated by performing an experiment with EAM heating and then performing an experiment without EAM heating, both in the same mean state, and subtracting the difference. The EAM-forced experiments apply a constant diabatic heating via Gaussian bubble with a maximum of 2 K day\(^{-1}\) centered at 30° N, 120° E and 300 hPa (see Supp. Fig. 1), similar to Zhu and Li (2016). Note that we also tested strong–weak EAM by running an experiment with positive heating forcing and an experiment with negative heating (cooling) in EAM location. Because patterns were similar, they have been excluded for the sake of brevity.

Finally, we address both indirect and direct ENSO modulation to the EAM-forced response as we did with the linear QG solutions: Indirect ENSO modulation was assessed by adjusting the observed and CAM5_obsSST mean state based on surface temperature climatology composited during El Niño or La Niña months, and direct ENSO modulation was assessed by running experiments with both ENSO-related forcing and EAM forcing and subtracting the original EAM-forced response. El Niño heating/La Niña cooling forcing location was chosen based on El Niño–La Niña divergence anomaly composites. A list and description of the experiments and their differences in purpose and setup can be found in Table 1.

### 2.5 EAM and ENSO definitions

The EAM index was defined by upper-level circulation as in Zhao et al. (2015): \(\text{Std}[U_{200}(2.5°–10° N, 105°–140° E) – U_{200}(17.5°–22.5° N, 105°–140° E) + U_{200}(30°–37.5° N, 105°–140° E)]\), where \(U_{200}\) is the 200-hPa zonal wind averaged within the domain outlined in the parentheses, and Std indicates a standardization of the timeseries of the index.

### Table 1 A list and description of experiments conducted with the dry nonlinear AGCM

| Experiment set | ENSO included? | Mean state | Experiments with respective forcing |
|----------------|----------------|------------|-----------------------------------|
| EAM-forced    | No             | Observations | 1. No forcing
|                |                | CAM5_climoSST | 2. EAM forcing
|                |                | CAM5_obsSST   | 1. No forcing
|                |                |              | 2. EAM forcing
| EAM-forced    | In mean state only | Observations during El Niño | 1. No forcing
| with indirect |                | Observations during La Niña | 2. EAM forcing
| ENSO          |                | CAM5_obsSST during El Niño | 1. No forcing
| modulation    |                |              | 2. EAM forcing
|                |                | CAM5_obsSST during La Niña | 1. No forcing
|                |                |              | 2. EAM forcing
| EAM-forced    | Yes            | Observations | 1. EAM forcing
| with direct   |                |              | 2. EAM and El Niño forcing
| ENSO          |                |              | 3. EAM and La Niña forcing
| modulation    |                | CAM5_obsSST   | 1. EAM forcing
|                |                |              | 2. EAM and El Niño forcing
|                |                |              | 3. EAM and La Niña forcing

Note that responses are calculated by subtracting the respective time-mean of Exp1 from the time-mean of Exp2 or Exp3 (i.e. Exp2 – Exp1 or Exp3 – Exp1). Statistical significance is calculated by dividing response by the 90-day moving mean standard deviation of the difference.
Strong and weak EAM events are determined by the upper and lower terciles of the index, respectively. We chose this index because it characterizes the distinct EAM and WNPM precipitation anomalies, captures multiple timescales of variability, and has been linked to ENSO (Zhao et al. 2015).

We categorized ENSO events using the Niño3.4, defined as $[T^*(5° S–5° N, 170°–120° W)]$, where $T^*$ denotes SST anomalies. Niño events were defined by a monthly Niño3.4 index greater or equal to 0.5 °C, and Niña events were defined by a monthly Niño3.4 index less than or equal to − 0.5 °C. Otherwise, the ENSO event was considered neutral.

3 The ANA teleconnection in observations and CAM5
3.1 EAM-related precipitation and circulation patterns

The ANA teleconnection is the bridge between EAM rainfall and CONUS rainfall. First, we established this relationship using difference composites, calculated by subtracting the averaged precipitation anomalies during weak EAM events from the averaged precipitation anomalies during strong EAM events and standardizing the difference. The strong–weak EAM precipitation standardized anomalies are seen in Fig. 1a, c. In observations (Fig. 1a), the EAM structure is established, with significant positive rainfall anomalies within the 25°–35° N latitudinal belt and stretched over the western and central North Pacific. Negative rainfall anomalies are located south and north of the EAM rainfall belt as a part of the full monsoonal circulation. Over CONUS, there are two regions that have statistically significant EAM-related precipitation anomalies: the west coast and Plains-Midwest regions (corresponding green and magenta boxes). To analyze further, we found the kernel density estimation (KDE) of the standardized precipitation anomalies at each month and grid point within the box for the strong EAM events and weak EAM events (Fig. 1d, g). The distribution of strong EAM anomalies (red) is shifted right from the weak EAM anomalies (blue), indicating that a strong EAM is associated with a higher probability of wet anomalies in these regions. The CONUS difference composite pattern, though not statistically significant everywhere, is analogous to the northwest-southeast rainfall dipole from Zhu and Li (2016).

Generally, the CAM5_climoSST standardized precipitation anomaly difference composite (Fig. 1b) is similar to the observed over the EAM region. Over CONUS, the associated precipitation anomalies over the west coast are positive, but anomalies over the Plains-Midwest are negative, unlike...
observations. The KDEs corroborate the result over the west coast i.e. there is a higher probability of wet anomalies during a strong EAM over the west coast, most apparent at the right tail (Fig. 1e). While significant at the 95% confidence level, there is no discernible shift via visual inspection in the KDEs over the Plains-Midwest (Fig. 1h). We assert that this is largely due to changes in the tails that are hard to detect visually.

The precipitation anomaly pattern is more comparable to the observations in the CAM5_obsSST standardized precipitation difference composite (Fig. 1c). The structure of the anomalies for the entire monsoonal system is better represented. The CONUS precipitation anomaly pattern is improved over the Plains-Midwest, though these wet anomalies also extend to the southern Plains. However, there are dry anomalies in the eastern North Pacific and southern west coast that are not present in observations. The KDEs for both the west coast and Plains-Midwest indicate a statistically significant distribution change in rainfall anomalies between strong and weak EAM, but it is virtually undetectable by eye (Fig. 1f, i). This may suggest that the EAM-CONUS rainfall relationship is more affected by local processes in the model. Overall, CAM5_obsSST is able to capture the rainfall pattern, and we attempt to address biases or discrepancies in the remainder of the study by interpreting Rossby wave responses.

On this timescale of interest, the anomalou s zonally-asymmetric component of the circulation, stationary eddies, controls the pattern of anomalous moisture transport and rainfall (Liu et al. 1998). Figure 1j–l shows the zonally-asymmetric component of the strong–weak EAM Z250 anomaly difference composite (Z250'). In observations (Fig. 1j), a strong EAM is associated with a wave train pattern oriented meridionally over East Asia and Siberia. An anomalous ridge stretches from the EAM region to the North American coast, an anomalous trough is found off the coast of Baja California/US southwest and across the Gulf of Mexico, and an anomalous ridge is centered over Quebec/Hudson Bay. This pattern is weak over North America, but it is generally in agreement with previous studies that showed that anomalous troughs over western US are important to identify southerly winds and moisture transport into CONUS, fueling precipitation events (Mallapkour and Villarini 2016; Malloy and Kirtman 2020).

Unlike observations, the Z250’ difference composite pattern from CAM5_climoSST has more of a zonal wave train structure from the EAM region to North America, and, over North America, the west–east pattern is opposite to observations (Fig. 1k). The Z250’ difference composite pattern from CAM5_obsSST has a noticeably improved representation of the teleconnection over North America (Fig. 1l). This also dynamically explains the better comparison of the standardized precipitation anomaly differences between observations and CAM5_obsSST from Fig. 1a–c. In general, including prescribed observed global SST variability in CAM5 improved the representation of the ANA pattern and its associated precipitation patterns over North America. Even in the EAM region, the local stationary wave pattern in the observations is not particularly well captured by the model, somewhat surprisingly, although arguably better with observed SST forcing.

### 3.2 EAM-related patterns separated by ENSO phase

On seasonal-to-interannual scales, ENSO has been found to have an influence on the EAM response (Ding et al. 2011); therefore, conditioning the EAM-related pattern on ENSO phases for observations and CAM5_obsSST may reveal different impacts downstream.

The observed conditioned strong–weak EAM composites of standardized precipitation anomalies and Z250’ validate that ENSO phase is important for understanding CONUS impacts. Figure 2 shows the classification of the difference composites from Fig. 1 into ENSO phases by finding the strong–weak EAM during El Niño (top row), neutral ENSO (middle row), and La Niña (bottom row) months. Over the Plains-Midwest, EAM-related precipitation anomalies during El Niño are more robust than during neutral ENSO or La Niña (Fig. 2a–c). The west coast anomalies are sensitive to ENSO phase i.e. there are only statistically significant wet anomalies during the neutral phase. The largest differences in Z250’ are over the eastern North Pacific and North America (Fig. 2j–l), with La Niña being associated with an anomalous ridge over most of Canada and the Pacific Northwest. By examining the KDEs (Fig. 2d–i) of the standardized precipitation anomalies, during La Niña, the Plains-Midwest has a greater probability of dry anomalies during a strong EAM, unlike during the other phases.

The weaknesses in the model capturing the observed EAM-CONUS teleconnections are amplified when stratifying by ENSO phase. As seen in Fig. 3, there are wet anomalies over the west coast and Plains-Midwest during neutral ENSO months (Fig. 3b), further established by the slight shift in distribution for strong EAM events (Fig. 3e, h). The west coast precipitation anomalies are contingent on ENSO phase in CAM5_obsSST: during El Niño, there is a north–south pattern of wet-dry anomalies and no significant difference in the KDEs between strong and weak EAM months (Fig. 3a, d), and, during La Niña, there is a greater probability of dry anomalies overall (Fig. 3c, f). The Z250’ patterns over North America are also dissimilar: during El Niño, the Z250’ difference composite is not statistically significant over North America (Fig. 3j), and, during La Niña, there is a cross-Pacific wave train response, with an anomalous trough over western North America and an anomalous ridge over eastern North America (Fig. 3l). This explains the
extreme positive precipitation anomalies in the composites and increased probability of positive precipitation anomalies over the Plains-Midwest seen in the KDE shifts (Fig. 3c, i).

Overall, the ENSO-modulated ANA teleconnection is not well represented in CAM5_obsSST, especially during La Niña.

Fig. 2  a–c Observed strong–weak EAM precipitation standardized anomaly composites during a El Niño conditions, b neutral ENSO conditions, and c La Niña conditions. Green and magenta boxes denote West Coast and Plains-Midwest domains, respectively. Stippling indicates anomalies significant at the 90% confidence level. Sample sizes for composites are also annotated in upper-left corner. d–i Kernel density estimators of the precipitation standardized anomalies for the grid points corresponding to the d–f West Coast and g–i Plains-Midwest domains for (red) strong EAM months and (blue) weak EAM months. Single asterisks on the upper-left corner of panel indicates a distribution change significant at the 95% confidence level, and double asterisks indicates significance at the 99% confidence level, based on a Wilcoxon rank-sum test. j–l Observed zonally-asymmetric component of strong–weak EAM Z250 composites during j El Niño conditions, k neutral ENSO conditions, and l La Niña conditions.

Fig. 3 Same as Fig. 2, but for CAM5_obsSST experiment.
We note that one limitation of this experimental design is that the observations and model include different time periods: 1979–2019 for observations and 1972–2001 for CAM5_obsSST. However, we performed the same analysis for the overlap period—1979–2001—and found that results overall were similar and only subtle differences exist (cf. Supp. Figs. 2 and 3). Therefore, the different time periods between model and observations does not explain the discrepancies.

The ANA teleconnection was better represented in CAM5_obsSST, the experiment with prescribed observed SST variability. Yet, the distinct patterns from the ENSO phases were unalike—even opposite to—the observations. This motivates a need to understand the EAM-forced and ENSO-influenced responses. What are the roles of the mean state and divergence forcing, and why does the model fail to capture the observed relationships?

4 The roles of mean state and forcing

4.1 Steady-state linear QG solution results

The linear QG analysis presented here is examining the extent to which the EAM-CONUS ANA teleconnection can be understood in this simple framework. Essentially, the extent to which the linear QG model captures the CONUS component of the teleconnection can be used to understand, at least in part, why CAM5 fails and further bolsters the simple interpretation of the observational results.

As described in Sect. 2, the steady-state linear QG solution to EAM forcing in the mid-latitudes depends on the divergence forcing (F) structure and a mean state (U), both of which are different between the observations, CAM5_climoSST experiment, and CAM5_obsSST experiment. Figure 4 summarizes the different linear QG Z250 responses and attempts to isolate the U and F “contribution” to the pattern. The observed response can be seen in Fig. 4a: when inputting U250 climatology as U and strong–weak EAM DIV250 anomaly composite over EAM region as F, the solution is a meridionally oriented Z250 pattern over the EAM region and a zonally oriented wave train over the North Pacific and North America. While not perfect, there are many similarities to the ANA pattern from Fig. 1j, particularly the response over the EAM region and generally over the eastern North Pacific and North America.

The CAM5_climoSST response (Fig. 4b)—obtained by inputting its separate U250 climatology and DIV250 anomaly composite—is much different. The mid-latitude response has many similarities to Fig. 1k, including the response from f CAM5_climoSST experiment and g CAM5_obsSST experiment. The corresponding U is overlaid in light gray with contours of 10 and 20 m s⁻¹. The corresponding F is overlaid, with thick green contours as −0.5 and −1 s⁻¹×10⁻⁶ and thick purple contours as 0.5 and 1 s⁻¹×10⁻⁶.
north of 30° N over EAM region and western North Pacific (100°–180° E) and the general ridge-trough pattern over North America, though out-of-phase by ½ a wavelength. Many aspects of the ANA teleconnection from CAM5_climoSST as well as its contrast to the observed ANA pattern can be described by the steady-state linear QG response to the EAM forcing. Interestingly, the CAM5_obsSST response (Fig. 4c) looks identical to the CAM5_climoSST response despite the stark differences over CONUS seen in Fig. 1k, l.

We next fixed F as the divergence anomalies from observations (Fig. 4d, e), keeping $\overline{U}$ as corresponding to CAM5_climoSST or CAM5_obsSST, which improves the ANA pattern overall. The response over the EAM region improves, and the response with the $\overline{U}$ from CAM5_obsSST (Fig. 4e) looks more like its EAM-related pattern from Fig. 1l, especially over the mid-latitude Pacific, Alaska, and southwest US. This suggests that F greatly influences the structure of the ANA pattern, and the F biases in CAM5 are limiting its representation of the pattern. In addition, the contrasts in the Z250 response via $\overline{U}$ differences between CAM5_climoSST and CAM5_obsSST are highlighted. Though subtle, the different $\overline{U}$ (jet stream climatologies) in these CAM5 experiments can perhaps explain some aspect of the different Z250 responses over the North Pacific.

We also fixed $\overline{U}$ as the U250 climatology from observations (Fig. 4f, g) and kept F as corresponding to CAM5_climoSST or CAM5_obsSST, confirming former findings. The structure of these responses looks similar to the CAM5 responses without the fixed $\overline{U}$ (Fig. 4b, c), and dissimilar to the observed response (Fig. 4a), further validating the importance of improving the EAM-related divergence forcing, F, to correct the ANA pattern and associated impacts over CONUS.

In brief, the linear QG solutions captured many aspects of the observed ANA teleconnection (cf. Fig. 4 vs. Fig. 1j–l). The misrepresentation of the ANA teleconnection in CAM5 is greatly influenced by the CAM5 biases in the EAM forcing (F). Differences between the two CAM5 simulations (CAM5_climoSST vs. CAM5_obsSST) are very subtle and not well explained by this linear QG framework.

To understand the indirect ENSO influence on the ANA pattern in observations and the CAM5_obsSST simulation, F was conditioned on ENSO phase before solving for the Z250 response, i.e. F was input as the strong–weak EAM DIV250 anomaly during El Niño or La Niña. For the observed Z250 EAM-forced responses during El Niño and La Niña, there is little difference (Fig. 5a, b). This is also true for the CAM5_obsSST responses (Fig. 5c, d). ENSO’s influence on the location/structure of EAM divergence forcing does not modulate the ANA pattern. Nevertheless, the disparities between the observed (Fig. 5a, b) and CAM5_obsSST (Fig. 5c,d) responses can still be described by the differences in the F location/structure, once again.

**Fig. 5** Linear steady-state QG solutions of Z250 with $\overline{U}$ and F from a, b observations and c, d CAM5_climoSST experiment. a, c Solutions when F is conditioned on El Niño, and b, d solutions when F is conditioned on La Niña. The corresponding $\overline{U}$ is overlaid in light gray with contours of 10 and 20 m s$^{-1}$. The corresponding F is overlaid, with thick green contours as $-0.5$ and $-1$ s$^{-1} \times 10^{-6}$ and thick purple contours as 0.5 and 1 s$^{-1} \times 10^{-6}$.
\( \bar{U} \) was also conditioned on ENSO phase, i.e. \( \bar{U} \) was taken as the U250 composite during El Niño or La Niña, and stark differences between the responses with El Niño-related \( \bar{U} \) and La Niña-related \( \bar{U} \) are discerned in Fig. 6. In observations, the Z250 response with El Niño-related \( \bar{U} \) (Fig. 6a) is more amplified and has an eastward-shifted structure. Interestingly, the canonical El Niño pattern (typically linked to winter) is observed over North America, with an anomalous trough over southern CONUS and an anomalous ridge north of it. In contrast, the response with the La Niña-related \( \bar{U} \) (Fig. 6b) has an opposite pattern, closer to the response not conditioned on ENSO in Fig. 4a.

The CAM5_obsSST solutions reveal the limitations of the model mean state during ENSO. The response with the El Niño-related \( \bar{U} \) (Fig. 6c), like the observed response, has a wave pattern over the North Pacific, with an anomalous trough over the eastern North Pacific and anomalous ridge over the high-latitudes of North America. The wavelength of this pattern is shorter with the \( \bar{U} \) from CAM5_obsSST, and it does not include the anomalous trough over southern CONUS, but there are many parallels. Conversely, the response with the La Niña-related \( \bar{U} \) (Fig. 6d) has very little resemblance to the La Niña-related response from observations.

In short, conditioning the \( \bar{U} \) on ENSO does seem to explain (1) the variability in the ANA pattern and (2) large model biases in CAM5_obsSST during La Niña months. However, we note that these responses are not comparable to Figs. 2j–l and 3j–l, particularly over CONUS.

Direct ENSO modulation is most apparent over CONUS, seen in Fig. 7. In observations, El Niño modulation (Fig. 7a), though shifted \( \frac{1}{2} \) wavelength from the EAM-forced response, leads to general amplification of the EAM-forced response over CONUS at 120° W and 80° W and weakening of the response in central North America (cf. Fig. 4a). La Niña modulation (Fig. 7b) leads to a weakening at 120° W and 80° W. Although the ENSO forcing in the tropical Pacific is much stronger in CAM5_obsSST than observations (purple and green contours), the CAM5_obsSST solutions for El Niño modulation (Fig. 7c) and La Niña modulation (Fig. 7d) are weaker. There is little difference between ENSO phases over CONUS, suggesting that CAM5_obsSST ENSO-related mean state differences are not substantial enough to impact indirect ENSO modulation in CAM5.

Despite the usefulness in the linear QG solutions and framework, there are aspects of the ANA teleconnection, such as the subtropical Pacific response, not captured. In addition, this framework does not address concerns about variability of this response; differences in mean state between observations and CAM5 may influence the robustness of the EAM-forced response. Finally, while we added direct ENSO-related divergence forcing to this model to assess direct modulation, it is possible that their interaction is nonlinear. For these reasons, we utilize the dry nonlinear AGCM.

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**Fig. 6** Same as Fig. 5, but when \( \bar{U} \) is conditioned on a, c El Niño and b, d La Niña.
4.2 Dry nonlinear AGCM experiment results

The dry, nonlinear AGCM can test the sensitivity of responses with different mean states while also determining the robustness of the EAM-forced circulation response. Robustness is measured by calculating the 90-day moving mean standard deviation (σ) of the response, and dividing the responses according to this moving mean standard deviation σ. The following results are from experiments where the heating forcing location was based on the observed EAM-related divergence anomalies (cf. Fig. 4a). Results for the EAM-forced experiments with the mean states from CAM5 when the heating forcing is shifted 10° eastward (based on Fig. 4b, c) are provided in Supp. Fig. 4, though the findings are similar.

When the observational climatology was input as the mean state, the Z250 response (Fig. 8a) includes a robust elongated anomalous ridge over the North Pacific and anomalous trough centered over the southwestern US, indicated by the $> + 1\sigma$ and $< - 1\sigma$ anomalies, respectively. By comparing to Fig. 1j, it is apparent that the dry atmospheric dynamics in this AGCM can effectively simulate the observed ANA teleconnection. Because this is a dry model, we can consider the time-mean DIV250 response (Fig. 8d) as a proxy for large-scale precipitation patterns associated with the heating forcing. The observed DIV250 response also corresponds with EAM-related patterns from Fig. 1a. There is upper-level divergence over the western coast, central US, and regions of Quebec, which are dynamically linked to the wet anomalies. Additionally, there is upper-level convergence over Alberta/Northwest Territories, southwest US, and the eastern coast, dynamically linked to dry anomalies. However, this response is relatively weak as σ values are low; this suggests that large-scale precipitation patterns due to EAM heating forcing are subject to significant variability on the seasonal-to-interannual timescales. In fact, differences/modulation of DIV250 responses are not discernible and not robust (Figs. 8e, f, 9, 10, 11 and 12c, d); therefore, we concentrate on Z250 responses/differences for remainder of results.

Over East Asia and North Pacific, the dry AGCM responses when the CAM5_climoSST and CAM5_obsSST experiment climatologies were input as the mean state are well represented (Fig. 8b, c). The Z250 response is a more zonal wave pattern, corresponding well to Fig. 1k, l. However, the strong anomalous trough centered over southern
US and northern Mexico in the dry AGCM is unlike the composite pattern in Fig. 1k, l. The DIV250 responses are not significantly different from the experiment with the mean state from observations. Overall, while the dry AGCM responses can explain CAM5 ANA patterns over the North Pacific well, there are discrepancies over North America between the dry AGCM responses and composite patterns that suggest additional processes are likely to be important. Other sources of forcing (e.g. ENSO) and/or land–atmosphere feedbacks may significantly contribute in simulating the ANA teleconnection in CAM5 experiments.

The dry AGCM response differences in Figs. 9 and 10 suggest that El Niño- or La Niña-related mean state differences alter the responses, confirming the steady-state linear QG solution results. The difference in dry AGCM $Z_{250}$ response with the observational El Niño-composited climatology as the mean state (Fig. 9a) has an anomalous ridge over high-latitude East Asia, an anomalous trough centered over the Bering Sea, an anomalous ridge over the Gulf of Alaska, and an anomalous trough over central North America. This is opposed to the response with the La Niña-composited climatology as the mean state (Fig. 10a), which reveals an opposite pattern. By comparing with the conditional composites from Fig. 2j, l, we find that the indirect modulation with the dry AGCM simulates ENSO-related patterns overall over North America, but not over East Asia and the North Pacific.

The responses with the CAM5_obsSST Niño- and Niña-composited climatologies (Figs. 9b and 10b) are very similar to each other, just as with the linear QG solutions. ENSO-related mean state differences are not strong in CAM5, reducing indirect ENSO modulation. By comparing with the conditional composites from Fig. 3j, l, we observe that the dry AGCM does not capture the ENSO-related patterns from CAM5 well overall, but there is somewhat more of a zonally oriented trough-ridge pattern over CONUS associated with the response with the La Niña-composited climatology, as in Fig. 3l. In general, indirect ENSO modulation may not be sufficient in describing the ENSO-related ANA patterns in this dry AGCM, especially over the North Pacific and for patterns from CAM5_obsSST.

Fig. 8 Dry AGCM EAM heating response, including the a–c asymmetric component of the time-mean $Z_{250}$ and d–f time-mean DIV250, for when the inputted mean state is from a, d observations, b, e CAM5_climoSST experiment, and c, f CAM5_obsSST experiment. The 1σ (solid black) and − 1σ (dashed black) values—are overlaid, representing ~67% of the variability of the response assuming Gaussian statistics. Heating forcing (purple contours) of 0.5, 1, and 1.5 K day$^{-1}$ are also overlaid on $Z_{250}$ responses.
We effectively extract direct ENSO modulation, seen in Figs. 11 and 12. Note that these should be interpreted as only the ENSO-modulated portion of the response, not the total EAM- and ENSO-forced response. The El Niño-modulated $Z_{250}$ with the observational mean state (Fig. 11a) shows a band of negative anomalies stretched over the subtropical Pacific, which would weaken the EAM-forced elongated North Pacific ridge and strengthen the trough over western CONUS (cf. Fig. 8a). In addition, there are positive anomalies over Alaska and central-eastern CONUS. Collectively, this results suggests that El Niño promotes strengthening of the west–east trough-ridge pattern, or constructive
Fig. 11 Dry AGCM El Niño-modulated part of the heating response (see Table 1 for details).  

- **a, b** Asymmetric component of the time-mean Z250 and **c, d** time-mean DIV250, for when the inputted mean state is taken from **a, c** observations and **b, d** CAM5_obsSST experiment. The 1σ (solid black) and − 1σ (dashed black) values are overlaid. Heating forcing (purple contours) of 0.5, 1, and 1.5 K day$^{-1}$ from combined EAM- and El Niño-forced experiment are also overlaid on Z250 responses.

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Fig. 12 Same as Fig. 11, but the La Niña-modulated part of the heating response. Heating forcing (purple contours) of − 0.5, − 1, and − 1.5 K day$^{-1}$ are included.
interference with the ANA pattern, over CONUS, influencing enhanced precipitation over the Plains-Midwest (cf. Fig. 2a). Conversely, the La Niña-modulated $Z_{250}$ with the observational mean state (Fig. 12a) shows positive anomalies in the subtropical Pacific and western CONUS. The pattern over North America is opposite to El Niño-modulated $Z_{250}$, suggesting destructive interference with the EAM-forced trough over western CONUS and explaining the decreased precipitation over the Plains-Midwest during La Niña (Fig. 2c).

El Niño- and La Niña-modulated $Z_{250}$ with the CAM5_obsSST mean state (Figs. 11b and 12b) have similar destructive and constructive interference of ANA pattern, respectively, over the subtropical Pacific, as in the results with the observational mean state. However, there is an anomalous ridge centered over CONUS in both the El Niño- and La Niña-modulated $Z_{250}$ patterns, not aligning with observations nor explaining the ENSO-related patterns from Fig. 3. Again, CONUS patterns in the dry AGCM with the CAM5_obsSST mean state are not simulated well, perhaps emphasizing the differences in the important processes behind the ANA pattern in observations and CAM5.

5 Summary and discussion

Summer CONUS rainfall remains a forecast challenge. Understanding rainfall variability and its links with quasi-persistent features, e.g. monsoons and SST anomalies, is imperative for long-range forecasting. In this study, we established the statistical and dynamical relationship between the EAM and CONUS precipitation and circulation anomaly patterns, the ANA teleconnection, which impacts the US west coast and Plains-Midwest regions (Fig. 1). This confirmed Zhu and Li (2016) study’s ANA rainfall pattern and the many studies that have linked Plains-Midwest climate to EAM activity (Lopez et al. 2019), but the link between the EAM and US west coast precipitation is relatively novel. We compared the ANA pattern from observations to its representation in a state-of-the-art atmospheric model, CAM5, where one experiment prescribed global climatological SSTs (CAM5_climoSST), and the other prescribed global observed SSTs (CAM5_obsSST). The ANA patterns from the CAM5_obsSST experiment were more similar to observations, suggesting that SST variability is important for simulating this teleconnection, although some notable difference between CAM5_obsSST and the observations were identified.

Because ENSO has an established influence on the EAM (Wang et al. 2001, 2008; Ding et al. 2011), we investigated the effect of ENSO’s modulation on the ANA teleconnection in observations and CAM5_obsSST (Figs. 2 and 3). The ANA patterns did depend on ENSO phase over the Plains-Midwest region. However, representation of the ENSO-modulated ANA patterns in CAM5_obsSST was limited; in fact, Niña-related ANA patterns over the Plains-Midwest between observations and CAM5_obsSST were of opposite sign (dry vs. wet anomalies, respectively).

Next, we used steady-state linear QG solution framework and a dry nonlinear AGCM to diagnose to what extent the observations and the CAM5 simulations can be understood in terms in this simplified framework as well as how much of the ANA teleconnection is due to nonlinear processes. We summarize the results from this section as follows:

1. CAM5 limitations in representing the ANA pattern are due to CAM5 biases in the EAM forcing location/structure (Figs. 4, 6 and 7).

2. Conditioning the mean state on ENSO generates distinctive ANA patterns. The biases in the ENSO-related mean state in CAM5_obsSST at least partly describes the erroneous representation of the La Niña-related ANA teleconnection in CAM5_obsSST (Fig. 6).

3. Nonlinear, dry atmospheric dynamics is essential for producing the EAM-forced pattern (cf. Figs. 1j and 8a), particularly over the subtropics.

4. Both indirect ENSO modulation and direct ENSO modulation are important for simulating details of the ANA pattern over CONUS (Figs. 7, 11 and 12). Generally, constructive (destructive) interference of El Niño (La Niña) patterns with the ANA pattern over CONUS explains enhanced (weakened) precipitation signals, particularly over the Plains-Midwest, which is reproduced well in both the linear QG solutions and nonlinear dry AGCM.

The forcing location biases in CAM5 are apparent (e.g. purple and green contours in Fig. 4a versus Fig. 4b, c). Many climate models fail to capture the variability and structure of the ASM systems. In general, models without air-sea coupling, such as CAM5, have a degraded SST-precipitation relationship over the EAM region (Islam et al. 2013), and biases in the West Pacific subtropical high have been linked to rainfall biases over this EAM rainfall belt (Liu et al. 2019); however, it is beyond the scope of this paper to diagnose reasons for this particular EAM divergence forcing bias in CAM5. In addition, these results complement previous studies that explored teleconnection sensitivity to mean state (Henderson et al. 2017; Kim et al. 2020; Wang et al. 2020; Li et al. 2021); model mean state, especially relating to jet stream strength and location, is essential for simulating correct Rossby wave responses and downstream impacts (O’Reilly et al. 2018; Li et al. 2021). In this study, mean state biases in CAM5 were important when considering the ENSO-modulated ANA pattern over North America.
Not all features of the ANA pattern nor the CAM5 biases were simulated/explained sufficiently with the linear QG model nor the dry AGCM (e.g., discrepancies over CONUS patterns between the dry AGCM responses with CAM5 mean states in Figs. 8, 9, 10, 11 and 12 and CAM5 ANA pattern over CONUS in Figs. 1 and 3). Neither model has land-moisture feedbacks, which are likely key in simulating the combined EAM and ENSO influence. Generally, summertime predictability has been linked to soil moisture and atmosphere-land processes (Dirmeyer et al. 2003; Koster et al. 2006; Schubert et al. 2008; Burgman and Jang 2015; Jong et al. 2021); therefore, there should be more analysis done on specific connections between ANA-related rainfall and atmosphere-land coupling.

Future studies should consider additional sources of SST variability besides ENSO. Previous studies have suggested that North Atlantic SST variability is vital for understanding CONUS hydroclimate (Schubert et al. 2008; Weaver et al. 2009; Malloy and Kirtman 2020). Extratropical Pacific SST variability (e.g. Pacific Decadal Oscillation) influences both ENSO events and mid-latitude Rossby wave propagation on the interdecadal timescale (Lee et al. 2011; Zhu et al. 2013; Burgman and Jang 2015; Song and Zhou 2015). Indian Ocean SST variability, including the Indian Ocean Dipole (IOD), interacts with both ENSO and ASM variability (Islam et al. 2013; Du et al. 2016; Lee and Wang 2016).

The study’s focus was on contemporary, monthly-mean ENSO modulation of the ANA teleconnection, but the transitional, slowly evolving aspects of ENSO-EAM links (e.g., post-peak Niño effects) might also be important (Weaver et al. 2016; Li et al. 2019). ENSO variability on the interdecadal timescale (Wu et al. 2009; Zhu and Li 2018) may change how it modulates the ANA pattern or robustness/predictability of responses.

Both complex AGCMs and simple dry AGCMs were employed to simulate processes needed to reproduce the ANA teleconnection, ENSO modulation, and their impact on CONUS circulation and rainfall. Whether or not the better understanding of the predictability of quasi-persistent teleconnections and Rossby wave responses translates to an increase in long-range rainfall forecast skill remains an open question.

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**Data availability** Enquiries about data availability should be directed to the authors.

**Declarations**

**Conflict of interest** The authors have not disclosed any competing interests.

**References**

Becker E, van den Dool H, Zhang Q (2014) Predictability and forecast skill in NMME. J Clim 27:5891–5906. https://doi.org/10.1175/JCLI-D-13-00597.1

Beverley JD, Woolnough SJ, Baker LH, Johnson SJ, Weisheimer A (2019) The northern hemisphere circumglobal teleconnection in a seasonal forecast model and its relationship to European summer forecast skill. Clim Dyn 52(5):3759–3771. https://doi.org/10.1007/s00382-018-4371-4

Beverley JD, Woolnough SJ, Baker LH, Johnson SJ, Weisheimer A, O’Reilly CH (2021) Dynamical mechanisms linking Indian monsoon precipitation and the circumglobal teleconnection. Clim Dyn 57(9):2615–2636. https://doi.org/10.1007/s00382-021-05825-6

Brenner S, Mitchell K (Kenneth Erwin), Yang C, U.S. Air Force Geophysics Laboratory, Atmospheric Sciences Division (1984) The AFGL global spectral model: expanded resolution baseline version. Hanscom AFB, Massachusetts: Air Force Geophysics Laboratory, Air Force Systems Command, United States Air Force. Retrieved from https://catalog.hathitrust.org/Record/102326036. Home. Accessed Feb 2021

Burgman RJ, Jang Y (2015) Simulated U.S. drought response to interannual and decadal Pacific SST variability. J Clim 28:4688–4705. https://doi.org/10.1175/JCLI-D-14-00247.1

Ciancarelli B, Castro CL, Woodhouse C, Dominguez F, Chang H, Carrillo C, Griffin D (2014) Dominant patterns of US warm season precipitation variability in a fine resolution observational record, with focus on the southwest. Int J Climatol 34(3):687–707. https://doi.org/10.1002/joc.3716

DeAngelis AM, Wang H, Koster RD, Schubert SD, Chang Y, Marshall J (2020) Prediction skill of the 2012 U.S. Great Plains flash drought in subseasonal experiment (SubX) models. J Clim 33(14):6229–6253. https://doi.org/10.1175/JCLI-D-19-0863.1

Ding Q, Wang B (2005) Circumglobal teleconnection in the northern hemisphere summer. J Clim 18:3483–3505. https://doi.org/10.1175/1155JCLI3473.1

Ding Q, Wang B, Wallace JM, Branstator G (2011) Tropical–extratropical connections in boreal summer: observed interannual variability. J Clim 24:1878–1896. https://doi.org/10.1175/2011JCLI3621.1

Dirmeyer PA, Fennessy MJ, Marx L (2003) Low skill in dynamical prediction of boreal summer climate: grounds for looking beyond sea surface temperature. J Clim 16:995–1002. https://doi.org/10.1175/1520-0442(2003)016%3C0995:LSIDPO%3E2.0.CO;2

Du Y, Li T, Xie Z et al (2016) Interannual variability of the Asian subtropical westerly jet in boreal summer and associated with circulation and SST anomalies. Clim Dyn 46:2673. https://doi.org/10.1007/s00382-015-2723-x

Ha KJ, Seo YW, Lee JY, Kripalani RH, Yun KS (2018) Linkages between the South and East Asian summer monsoons: a review
The summer Asia–North America teleconnection and its modulation by ENSO in Community Atmosphere…

and revisit. Clim Dyn 51(11):4207–4227. https://doi.org/10.1007/s00382-017-3773-z

Hao Z, Singh VP, Xia Y (2018) Seasonal drought prediction: challenges, challenges, and future prospects. Rev Geophys 56:108–141. https://doi.org/10.1002/2016RG000549

Henderson SA, Maloney ED, Son S (2017) Madden–Julian oscillation
The summer Asia–North America teleconnection and its modulation by ENSO in Community Atmosphere…

Hu Q, Feng S (2012) AMO- and ENSO-driven summertime circulation

Hersbach H, Bell B, Berrisford P, Hirahara S, Horányi A, Muñoz-

Hu ZZ, Kumar A, Jha B, Huang B (2019) An update on the estimate of pre-

Kim D, Lee S, Lopez H, Goes M (2020) Pacific mean-state control

Kirtman BP, Paolino DA, Kinter JL III, Straus DM (2001) Impact of

Jong B, Ting M, Seager R (2021) Assessing ENSO summer telecon-

Jha B, Kumar A, Hu ZZ (2019) An update on the estimate of pre-

Islam S, Tang Y (2017) Simulation of different types of ENSO impacts
 on South Asian Monsoon in CCSM4. Clim Dyn 48(3–4):893–911. https://doi.org/10.1007/s00382-016-3117-4

Islam S, Tang Y, Jackson PL (2013) Asian monsoon simulations by
Community Climate Models CAM4 and CCSM4. Clim Dyn 41(9–10):2617–2642. https://doi.org/10.1007/s00382-013-1752-6

Jha B, Kumar A, Hu ZZ (2019) An update on the estimate of pre-
dictability of seasonal mean atmospheric variability using North
American Multi-Model Ensemble. Clim Dyn 53(12):7397–7409. https://doi.org/10.1007/s00382-016-3217-1

Jong B, Ting M, Seager R (2021) Assessing ENSO summer telecon-
nexions, and predictability in North America. J Clim 34(9):3629–3643. https://doi.org/10.1175/JCLI-D-20-0761.1

Kim D, Lee S, Lopez H, Goes M (2020) Pacific mean-state control of Atlantic multidecadal oscillation—El Niño relationship. J Clim 33(10):4273–4291. https://doi.org/10.1175/JCLI-D-19-0398.1

Kirtman BP, Paolino DA, Kinter JL III, Strauss DM (2001) Impact of tropical seasonal SST variability on seasonal mean climate. Mon Weather Rev 129:853–868. https://doi.org/10.1175/1520-0493(2001)129<0853:OTTSSV>2.0.CO;2

Koster RD, Sud YC, Guo Z, Dirmeyer PA, Bonan G, Oleson KW et al (2006) GLACE: the global land–atmosphere coupling experiment. Part I: overview. J Hydrometeorol 7(4):590–610. https://doi.org/10.1175/JHM510.1

Lau W, Weng H (2002) Recurrent teleconnection patterns linking summertime precipitation variability over East Asia and North America. J Meteorol Soc Jpn 80:1309–1324. https://doi.org/10.2151/jmsj.80.1309

Lau K-M, Kim K-M, Lee J-Y (2004) Interannual variability, global teleconnection and potential predictability associated with the Asian summer monsoon. In: Chang CP (ed) East Asian monsoon. World Scientific, Singapore. https://doi.org/10.1142/9789812701411_0004

Lee S, Wang B (2016) Regional boreal summer intraseasonal oscillation over Indian Ocean and Western Pacific: comparison and predictability study. Clim Dyn 46:2213–2229. https://doi.org/10.1007/s00382-015-2698-7

Lee S, Wang C, Mapes BE (2009) A simple atmospheric model of the local and teleconnection responses to tropical heating anomalies. J Clim 22(2):272–284. https://doi.org/10.1175/2008JCLI2303.1

Lee JY, Wang B, Ding Q, Ha KJ, Ahn JB, Kumar A et al (2011) How predictable is the Northern Hemisphere summer upper-tropospheric circulation? Clim Dyn 37(5):1189–1203. https://doi.org/10.1007/s00382-010-0909-9

Li J, Ding R, Wu Z, Zhong Q, Li B, Li J (2019) Inter-decadal change in potential predictability of the East Asian summer monsoon. Theor Appl Climatol 136(1):403–415. https://doi.org/10.1007/s00704-018-2482-9

Li KKK, Tam CY, Lau NC, Sohn S-J, Ahn JB, O’Reilly C (2021) Forcing mechanism of the Silk Road pattern and the sensitivity of Rossby-wave source hotspots to mean-state winds. Q J R Meteorol Soc 47:2533–2546. https://doi.org/10.1002/qj.4039

Liu AZ, Ting M, Wang H (1998) Maintenance of circulation anomalies during the 1988 drought and 1993 floods over the United States. J Atmos Sci 55:2810–2832. https://doi.org/10.1175/1520-0469(1998)055<2810:MOCADT>2.0.CO;2

Liu F, Li T, Wang H, Deng L, Zhang Y (2016) Modulation of boreal summer intraseasonal oscillations over the Western North Pacific by ENSO. J Clim 29(20):7189–7201. https://doi.org/10.1175/JCLI-D-15-0831.1

Liu Y, Ke Z, Ding Y (2019) Predictability of East Asian summer monsoon in seasonal climate forecast models. Int J Climatol 39(15):5688–5701. https://doi.org/10.1002/joc.6180

Lopez H, Lee S-K, Dong S, Goni G, Kirtman B, Atlas R, Kumar A (2019) East Asian Monsoon as a modulator of U.S. Great Plains heatwaves. J Geophys Res Atmos 124:6342–6358. https://doi.org/10.1029/2018JD030151

Mallakpour I, Villarini G (2016) Investigating the relationship between the frequency of flooding over the central United States and large-scale climate. Adv Water Resour 92:159–171. https://doi.org/10.1016/j.adwres.2016.04.008

Malloy KM, Kirtman BP (2020) Predictability of midsummer Great Plains low-level jet and associated precipitation. Weather Forecast 35(1):215–235. https://doi.org/10.1175/WAF-D-19-0103.1

Mariotti A, Baggett C, Barnes EA, Becker E, Butler A, Collins DC, Dirmeyer PA, Ferranti L, Johnson NC, Jones J, Kirtman BP, Lang AL, Molod A, Newman M, Robertson AW, Schubert S, Waliser DE, Albers J (2021) Windows of opportunity for skillful forecasts of seasonal to subseasonal and beyond. Bull Am Meteorol Soc 101(5):E608–E625. https://doi.org/10.1175/BAMS-D-18-0326.1

Moon JY, Wang B, Ha KJ, Lee JY (2013) Teleconnections associated with Northern Hemisphere summer monsoon intraseasonal oscillation. Clim Dyn 40(11–12):2761–2774. https://doi.org/10.1007/s00382-012-1394-0

Neale RB et al (2010) Description of the NCAR Community Atmosphere Model (CAM 5.0). NCAR technical note NCAR/TN-486+STR, National Center for Atmospheric Research, Boulder

O’Reilly CH, Woollings T, Zanna L, Weisheimer A (2018) The impact of tropical precipitation on summertime Euro-Atlantic circulation via a circumglobal wave train. J Clim 31:6481–6504. https://doi.org/10.1175/JCLI-D-17-0451.1

Pegion PJ, Kumar A (2010) Multimodel estimates of atmospheric response to modes of SST variability and implications for droughts. J Clim 23:4327–4341. https://doi.org/10.1175/2010JCLI3295.1

Schubert SD, Suarez MJ, Pegion PJ, Kistler MA, Kumar A (2002) Predictability of zonal means during boreal summer. J Clim 15:420–434. https://doi.org/10.1175/1520-0442(2002)015<0420:POZMIDB>2.0.CO;2

Schubert SD, Suarez MJ, Pegion PJ, Koster RD, Bacmeister JT (2008) Potential predictability of long-term drought and pluvial conditions in the U.S. Great Plains. J Clim 21(4):802–816. https://doi.org/10.1175/2007JCLI1741.1

Schubert S, Wang H, Suarez M (2011) Warm season subseasonal variability and climate extremes in the Northern Hemisphere: the role of stationary Rossby waves. J Clim 24:4773–4792. https://doi.org/10.1175/JCLI-D-10-05035.1
Sheffield J, Camargo SJ, Fu R, Hu Q, Jiang X, Johnson N, Karnauskas KB, Kim ST, Kinter J, Kumar S, Langenbrunner B, Maloney E, Mariotti A, Meyerson JE, Neelin JD, Nigam S, Pan Z, Ruiz-Barradas A, Seager R, Serra YL, Sun D, Wang C, Xie S, Yu J, Zhang T, Zhao M (2013) North American climate in CMIP5 experiments. Part II: evaluation of historical simulations of intraseasonal to decadal variability. J Clim 26:9247–9290. https://doi.org/10.1175/JCLI-D-12-00593.1

Slater LJ, Villarini G, Bradley AA (2016) Evaluation of the skill of North-American Multi-Model Ensemble (NMME) global climate models in predicting average and extreme precipitation and temperature over the continental USA. Clim Dyn. https://doi.org/10.1007/s00382-016-3286-1

Song F, Zhou T (2015) The crucial role of internal variability in modulating the decadal variation of the East Asian summer monsoon–ENSO relationship during the twentieth century. J Clim 28(18):7093–7107. https://doi.org/10.1175/JCLI-D-14-00783.1

Sperber KR, Annamalai H, Kang IS, Kitoh A, Moise A, Turner A et al (2013) The Asian summer monsoon: an intercomparison of CMIP5 vs. CMIP3 simulations of the late 20th century. Clim Dyn 41(9–10):2711–2744. https://doi.org/10.1007/s00382-012-1607-6

Tian D, Wood EF, Yuan X (2017) CFSv2-based sub-seasonal precipitation and temperature forecast skill over the contiguous United States. Hydrol Earth Syst Sci 21(3):1477–1490. https://doi.org/10.5194/hess-21-1477-2017

Ting M, Wang H (1997) Summertime U.S. precipitation variability and its relation to Pacific sea surface temperature. J Clim 10:1853–1873. https://doi.org/10.1175/1520-0442(1997)010%3c1853:SUPTIV%3e2.0.CO;2

Trenberth KE, Branstator GW, Karoly D, Kumar A, Lau NC, Ropelewski C (1998) Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperatures. J Geophys Res 103(C7):14291–14324. https://doi.org/10.1029/97JC01444

Wang B, Wu R, Lau K (2001) Interannual variability of the Asian summer monsoon: contrasts between the Indian and the Western North Pacific–East Asian monsoons. J Clim 14:4073–4090. https://doi.org/10.1175/1520-0442(2001)014%3c4073:IVOTAS%3e2.0.CO;2

Wang B, Wu Z, Li J, Liu J, Chang CP, Ding Y, Wu G (2008) How to measure the strength of the East Asian summer monsoon. J Clim 21(17):4449–4463. https://doi.org/10.1175/2008JCLI1831.1

Wang J, Kim H, Kim D, Henderson SA, Stan C, Maloney ED (2020) MJO teleconnections over the PNA region in climate models. Part II: impacts of the MJO and basic state. J Clim 33(12):5081–5101. https://doi.org/10.1175/JCLI-D-19-0865.1

Weaver SJ, Schubert S, Wang H (2009) Warm season variations in the low-level circulation and precipitation over the Central United States in observations, AMIP simulations, and idealized SST experiments. J Clim 22:5401–5420. https://doi.org/10.1175/2009JCLI2984.1

Weaver SJ, Baxter S, Harnos K (2016) Regional changes in the interannual variability of U.S. warm season precipitation. J Clim 29:5157–5173. https://doi.org/10.1175/JCLI-D-14-00803.1

Wu B, Zhou T, Li T (2009) Contrast of rainfall–SST relationships in the Western North Pacific between the ENSO-developing and ENSO-decaying summers. J Clim 22(16):4398–4405. https://doi.org/10.1175/2009JCLI2648.1

Xie P, Arkin PA (1997) Global precipitation: a 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. Bull Am Meteorol Soc 78(11):2539–2558. https://doi.org/10.1175/1520-0477(1997)078%3c2539:GPAYMA%3e2.0.CO;2

Yang Y, Zhu Z, Li T, Yao M (2020) Effects of Western Pacific intraseasonal convection on surface air temperature anomalies over North America. Int J Climatol 40(6):2913–2923. https://doi.org/10.1002/joc.6373

Zhao G, Huang G, Wu R, Tao W, Gong H, Qu X, Hu K (2015) A new upper-level circulation index for the East Asian summer monsoon variability. J Clim 28(24):9977–9996. https://doi.org/10.1175/JCLI-D-15-0272.1

Zhao S, Deng Y, Black RX (2018) An intraseasonal mode of atmospheric variability relevant to the U.S. hydroclimate in boreal summer: dynamic origin and East Asia connection. J Clim 31:9855–9868. https://doi.org/10.1175/JCLI-D-18-0206.1

Zhou S, L’Heureux M, Weaver S et al (2012) A composite study of the MJO influence on the surface air temperature and precipitation over the Continental United States. Clim Dyn 38:1459–1471. https://doi.org/10.1007/s00382-011-1001-9

Zhou F, Ren HL, Hu ZZ, Liu MH, Wu J, Liu CZ (2020) Seasonal predictability of primary East Asian summer circulation patterns by three operational climate prediction models. Q J R Meteorol Soc 146(727):629–646. https://doi.org/10.1002/qj.3697

Zhu Z, Li T (2016) A new paradigm for continental U.S. summer rainfall variability: Asia–North America teleconnection. J Clim 29:7313–7327. https://doi.org/10.1175/JCLI-D-16-0137.1

Zhu Z, Li T (2018) Amplified contiguous United States summer rainfall variability induced by East Asian monsoon interdecadal change. Clim Dyn 50:3523. https://doi.org/10.1007/s00382-017-3821-8

Zhu J, Huang B, Hu ZZ, Kinter JL, Marx L (2013) Predicting US summer precipitation using NCEP Climate Forecast System version 2 initialized by multiple ocean analyses. Clim Dyn 41(7–8):1941–1954. https://doi.org/10.1007/s00382-013-1785-x

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