The Forcing of Southwestern Asia Teleconnections by Low-Frequency Sea Surface Temperature Variability during Boreal Winter

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ABSTRACT

Southwestern Asia, defined here as the domain bounded by 20°–40°N and 40°–70°E, which includes the nations of Iraq, Iran, Afghanistan, and Pakistan, is a water-stressed and semiarid region that receives roughly 75% of its annual rainfall during November–April. The November–April climate of southwestern Asia is strongly influenced by tropical Indo-Pacific variability on intraseasonal and interannual time scales, much of which can be attributed to sea surface temperature (SST) variations. The influences of lower-frequency SST variability on southwestern Asia climate during November–April Pacific decadal SST (PDSST) variability and the long-term trend in SST (LTSST) is examined. The U.S. Climate Variability and Predictability Program (CLIVAR) Drought Working Group forced global atmospheric climate models with PDSST and LTSST patterns, identified using empirical orthogonal functions, to show the steady atmospheric response to these modes of decadal to multidecadal SST variability. During November–April, LTSST forces an anticyclone over southwestern Asia, which results in reduced precipitation and increases in surface temperature. The precipitation and tropospheric circulation influences of LTSST are corroborated by independent observed precipitation and circulation datasets during 1901–2004. The decadal variations of southwestern Asia precipitation may be forced by PDSST variability, with two of the three models indicating that the cold phase of PDSST forces an anticyclone and precipitation reductions. However, there are intermodel circulation variations to PDSST that influence subregional precipitation patterns over the Middle East, southwestern Asia, and subtropical Asia. Changes in wintertime temperature and precipitation over southwestern Asia forced by LTSST and PDSST imply important changes to the land surface hydrology during the spring and summer.

1. Introduction

The climate of southwestern Asia, defined here as the domain bounded by 20°–40°N, 40°–70°E, which contains the nations of Iraq, Iran, Afghanistan, and Pakistan, is very sensitive to variations of the tropical Indo-Pacific during its primary precipitation season of November–April. Southwestern Asia is strongly influenced by tropical Indo-Pacific climate variability on intraseasonal (Hoell et al. 2012) and interannual (Hoell et al. 2013) time scales by the Madden–Julian oscillation (MJO) (Barlow et al. 2005; Nazemosadat and Ghaedamini 2010; Barlow 2011) and El Niño–Southern Oscillation (ENSO) (Mariotti 2007; Hoell et al. 2014a, b), respectively. Recent works have implied that Indo-Pacific climate variability on decadal and longer time scales may also influence southwestern Asia (Meehl and Hu 2006; Lyon et al. 2014). Here, we examine the influences of lower-frequency Indo-Pacific sea surface temperature (SST) variability on
southwestern Asia climate during November–April. Specifically, we focus on the influences of Pacific decadal SST (PDSST) variability and the long-term trend in SST (LTSST).

Southwestern Asia is a water-scarce region (Oki and Kanae 2006) that receives roughly 75% of its annual precipitation during November–April. Rain-fed subsistence farming is widespread throughout southwestern Asia (e.g., Ramankutty et al. 2008; Ryan et al. 2012), so the combined effects of water scarcity and drought increase the potential for loss of life and property (Agrawala et al. 2001). Since 1999, southwestern Asia has endured two very notable droughts, in 1999–2001 (Barlow et al. 2002; Hoerling and Kumar 2003) and 2007–08 (Hoell et al. 2012), embedded within a decade of dry conditions (Hoell and Funk 2014; Lyon et al. 2014). The clustering of intense droughts within a decade of dry conditions has strong societal impacts and raises the issue of whether southwestern Asia is susceptible to regular decadal drought variations in addition to year-to-year droughts.

The atmospheric responses over southwestern Asia to intraseasonal and interannual tropical Indo-Pacific climate variability are very similar. Anomalous tropical convection over the Maritime Continent and anomalous tropical subsidence over the central tropical Pacific Ocean associated with the MJO (Barlow et al. 2005; Nazemosadat and Ghaedamini 2010; Barlow 2011) and ENSO (Hoell et al. 2014a,b) excite baroclinic Rossby waves over southwestern Asia (Barlow et al. 2007). The baroclinic Rossby waves over southwestern Asia interact with the mean climate, resulting in thermodynamically forced vertical motions and modifications to the regional moisture budget (Mariotti 2007; Hoell et al. 2014a,b). In the case of ENSO, the regional circulations and precipitation changes are intensified when the west Pacific Ocean is warmer than average (Hoell and Funk 2013; Hoell et al. 2014a,b). Furthermore, ENSO may also excite barotropic Rossby waves over an eastward-propagating teleconnection mechanism (Shaman and Tziperman 2005; Hoell et al. 2013).

Since 1999, west Pacific SSTs have continued to warm (e.g., Solomon and Newman 2012) and decadal variations of Pacific SST have caused for a cooling of the east Pacific (Lyon et al. 2014). These decadal to multidecadal changes in SST have occurred concomitantly with recent La Niña events characterized by central Pacific cooling and a warm west Pacific Ocean (Johnson 2013). These La Niña events resulted in several droughts across the globe (Hoerling and Kumar 2003) with particular severity over southwestern Asia (Barlow et al. 2002; Hoell et al. 2014b). Therefore, in this work, we isolate the influences of PDSST and LTSST on southwestern Asia climate from the previously investigated interannual SST variability associated with ENSO.

The U.S. Climate Variability and Predictability Program (CLIVAR) Drought Working Group, hereafter referred to as DWG, identified the SST patterns associated with LTSST and PDSST and forced atmospheric global climate models with these characteristic SST patterns. Here, we examine those model simulations in addition to observed data to assess how LTSST and PDSST influence the southwestern Asia climate during November–April. In section 2, we discuss the forced SST experiments and data used in this study. In section 3, we describe observed southwestern Asia climate variability. In sections 4 and 5, we examine the influence of SST patterns consistent with LTSST and PDSST on the November–April southwestern Asia climate, respectively. In section 6, we provide a summary.

2. Data and methods

a. Forced SST experiments

The DWG completed a suite of global climate model simulations forced by idealized SST anomaly patterns representative of important modes of climate variability. The primary goals of these experiments were to better understand the impacts of important modes of SST variability on atmospheric circulations and the role of land–atmosphere feedbacks on drought. A comprehensive summary of these experiments and the SST patterns utilized in the experiments can be found in Schubert et al. (2009) and Pegion and Kumar (2010). Here, we utilize two sets of idealized experiments across three atmospheric models to examine the impacts of PDSST and LTSST on southwestern Asia climate. The PDSST and LTSST experiments and the SST patterns used to drive those experiments are discussed below.

1) IDENTIFICATION OF TREND AND DEcadAL SST PATTERNS

The DWG isolated the primary patterns of global SST variability on interannual and longer time scales using rotated empirical orthogonal functions (EOFs) of annual averages of the 1901–2004 monthly SST dataset of Rayner et al. (2003). The EOFs were rotated using varimax rotation (Kaiser 1958) to better separate the leading patterns of Atlantic and Pacific SST variability, annual averages were used to highlight interannual and longer time scales, and the EOFs were calculated using only ice-free grid points to avoid contamination (Schubert et al. 2009; Pegion and Kumar 2010). The first EOF (LTSST; Fig. 1a) and associated principal component (PC; Fig. 1b), which explains 27.2% of the variance, describe the global trend in SST. The LTSST loads strongly in the Indo-Pacific basin and closely resembles the Pacific-only trend patterns in the absence of ENSO, highlighted by neutral
or cooling SST over the central tropical Pacific and warming west Pacific warm pool (Cane et al. 1997; Compo and Sardeshmukh 2010; Solomon and Newman 2012). The temporal variability of LTSST has demonstrated a nearly linear increase since the early 1960s. We examine atmospheric models forced by the LTSST pattern to investigate the influence of the global SST trend on southwestern Asia.

The DWG also identified a low-frequency decadal variability-only pattern over the Pacific. The low-frequency SST variability was isolated by filtering the global monthly SST to retain time scales of about six years and longer (Zhang et al. 1997). The first rotated EOF of the low-frequency SST describes LTSST, similar to Fig. 1a, and no experiments were conducted using this pattern. The second rotated EOF of the low-frequency SST (PDSST; Fig. 1c) and associated PC (Fig. 1d) describe the pan-Pacific decadal pattern of variability. We examine atmospheric models forced by the PDSST pattern to investigate the influence of pan-Pacific decadal variability (e.g., Barlow et al. 2001) on southwestern Asia.

2) EXPERIMENTAL DESIGN

As part of the DWG, three modeling centers completed global climate model experiments forced by both the LTSST (Fig. 1a) and the PDSST patterns (Fig. 1e). The participating modeling centers and the atmospheric models used are as follows: the National Aeronautics and Space Administration Goddard Space Flight Center (NASA GSFC) utilized the NASA Seasonal-to-Interannual Prediction Program, version 1 (NSIPP-1), atmospheric global climate model (Bacmeister et al. 1994; Schubert et al. 2004); the Geophysical Fluid Dynamics Laboratory (GFDL) utilized version 2.1 of the GFDL Atmospheric Model (AM2.1) (Delworth et al. 2006; Anderson et al. 2004; Milly and Shmakin 2002); and the National Centers for Environmental Prediction Climate Prediction Center (NCEP CPC) utilized the NCEP Global Forecast System (GFS) (Campana and Caplan 2005). Please refer to Schubert et al. (2009) and the aforementioned references for further description of each model.

The full SST patterns of LTSST and PDSST used to force the atmospheric global climate models were constructed by adding scaled versions of their characteristic patterns identified by EOF analysis to the monthly varying 1901–2004 SST climatology. The forcing patterns are fixed in time and do not have an annual cycle; however, the full SST forcing patterns do have an annual

![Fig. 1. (a) The leading rotated EOF of annual average 1901–2004 global SST anomaly and (b) associated PC, displaying the LTSST. Warm LTSST (LTw) and cool LTSST (LTc) thresholds, taken to occur at plus or minus one std dev from the mean, are shown by dashed lines in (b). (c) The second rotated EOF of low-pass-filtered (time scales greater than 6 yr) monthly 1901–2004 global SST anomaly and (d) associated PC, displaying the PDSST. Warm PDSST (PDw) and cool PDSST (PDC) thresholds, taken to occur at plus or minus one std dev from the mean, are shown by dashed lines in (d). The product of the each PC and associated EOF yields units of degrees Celsius.](image-url)
cycle as a result of varying SST climatology. The scaling factor of the PDSST EOF was plus or minus two standard deviations of the associated PC while the scaling factor of the LTSST was plus or minus one standard deviation of the associated PC. The PDSST EOF was scaled to highlight the atmospheric response to this SST pattern while the LTSST was scaled to highlight the 40-yr averages of the trend anomaly at the beginning and end of the 1901–2004 period. Two simulations were performed for the PDSST and the LTSST experiments corresponding to the warm and cool phase of each SST mode. It is important to note that the LTSST in Fig. 1a corresponds to the warm phase whereas the PDSST in Fig. 1c corresponds to the cold phase. Here, it assumed that the response is linear for warm and cold phases, and that the amplitude of the response for each phase is roughly half of the value of this difference. Control runs for each model forced by the monthly varying climatology were also analyzed. The NSIPP-1 and GFDL AM2.1 simulations were integrated for 50 years for all experiments while the GFS simulations were integrated for 35 years.

3) Analysis and statistical significance

The monthly climatic response to PDSST is taken to be the difference between the PDSST cooling experiment and PDSST warming experiment in each model, also referred to as PDc-PDw. The monthly climatic response to the LTSST pattern is taken to be the difference between the LTSST warming experiment and LTSST cooling experiments in each model, also referred to as LTw-LTc. Here, we examine the average response in the GFS, NSIPP-1, and GFDL AM2.1 during the November–April season.

A two-tailed Student’s t test is used to assess whether the November–April means of the control run are from the same population as the PDc-PDw and LTw-LTc in each model. Significance at $p < 0.10$ ($p < 0.05$ at each tail) is denoted by stippling.

b. Observational analysis

1) Data

Precipitation for 1901–2004 was drawn from the land-only Global Precipitation Climatology Centre (GPCC) monthly precipitation dataset version 6 (Becker et al. 2013) on a fixed $0.5^\circ \times 0.5^\circ$ latitude–longitude grid.

Monthly 200-hPa geopotential height for 1901–2004 was drawn from the Twentieth Century Reanalysis, version 2 (20CR2; Compo et al. 2011), on a fixed $2.0^\circ \times 2.0^\circ$ latitude–longitude grid. The monthly 200-hPa geopotential height from 20CR2 was compared with and shows strong similarities to the monthly 200-hPa geopotential height NCEP–National Center for Atmospheric Research (NCAR) Reanalysis 1 fields (Kalnay et al. 1996) on a fixed $2.5^\circ \times 2.5^\circ$ latitude–longitude pressure level grid beginning in 1948.

Surface temperature for 1901–2004 was drawn from the land-only University of Delaware, version 3.01, dataset on a fixed $0.5^\circ \times 0.5^\circ$ latitude–longitude grid. Please see http://climate.geog.udel.edu/~climate/html_pages/Global2011/README.GlobalTsT2011.html for documentation.

2) Analysis and statistical significance

The observed precipitation and geopotential height response to LTw-LTc and PDc-PDw during November–April were calculated through composites based upon thresholds of plus or minus one standard deviation of the LTSST PC (Fig. 1b) and PDSST PC (Fig. 1d), respectively, for the 1901–2004 period. For example, precipitation and geopotential height composites of PDc-PDw were constructed by calculating the difference between the average November–April response during PDc (plus one standard deviation from the PDSST PC mean) and the average response during the PDw (minus one standard deviation of the PDSST PC mean) during 1901–2004. PDc corresponds to positive values of the PDSST PC (Fig. 1d) as to highlight cooling of the central Pacific Ocean given the sign of the loadings in PDSST EOF (Fig. 1c). Statistical significance was assessed using a Monte Carlo approach. Randomly selected November–April seasons corresponding to the number of PDw occurrences were averaged and subtracted from randomly selected November–April seasons corresponding to the number of PDc occurrences. This process was repeated to generate 10000 composites. The 10000 composites were sorted according to rank at all grid points and statistical significance was computed at $p < 0.05$ for each tail of the distribution.

3. Observed southwestern Asia variability

The complex topography of southwestern Asia (Fig. 2a) strongly influences regional precipitation variations. The largest precipitation amounts fall on the western slopes of the region’s high mountains, notably over the Zagros Mountains of Iran and Iraq (Alijani 2008) and the western edge of the Tibetan Plateau. Despite regional precipitation maximums near the region’s high mountains, there are many deserts throughout southwestern Asia, specifically, the Iranian salt deserts, the Registan desert of Afghanistan, and the Karakum desert of Turkmenistan. The majority of annual precipitation over southwestern Asia falls during November–April (Fig. 2c) resulting from eastward moving storm systems (Martyn 1992) that are guided by the mean upper tropospheric west-to-east flow (Krishnamurti 1961; Schemann et al. 2009). We limit our analysis of southwestern Asia climate
Despite southwestern Asia’s propensity for drought and marked subregional variations in precipitation, a thorough historical and spatial analysis of precipitation is hampered by a sparse observation network in space and time. Figure 2b demonstrates that there were few monthly precipitation observations over southwestern Asia used in the GPCC precipitation dataset prior to the mid-twentieth century. The poor distribution of precipitation observation locations over southwestern Asia is highlighted in Fig. 3, which leaves uncertainty in Iraq, Afghanistan, and Pakistan. Therefore, observed precipitation throughout southwestern Asia during the twentieth century must be interpreted with caution on fine spatial and time scales. On the regional scale, gridded datasets such as GPCC have the capacity to qualitatively identify wet and dry years over southwestern Asia (Schiemann et al. 2008).

Temporal variations of November–April precipitation over southwestern Asia are assessed in Fig. 4 for the period of 1901–2004. The year-to-year changes in rainfall can be large, varying by as much as 60 mm from the seasonal mean of 160 mm. Year-to-year changes in November–April precipitation over southwestern Asia are a result of atmospheric teleconnections forced by processes over the Indian Ocean (Hoell et al. 2012, 2013), ENSO (Barlow et al. 2002; Hoell et al. 2014a,b), and the
November–April precipitation over southwestern Asia also varies on decadal and multidecadal time scales. The 10-yr running averages of southwestern Asia precipitation (black line in Fig. 4) indicate that decadal variations can be large during November–April relative to the long-term mean. Severe droughts over southwestern Asia have occurred during the most recent decadal shift during the late 1990s, concomitant with the cold phase of Pacific decadal SST variability (Lyon et al. 2014), during La Niña events (Barlow et al. 2002; Hoell et al. 2014a,b).

In light of decadal and multidecadal variability of southwestern Asia precipitation, we examine the observed precipitation patterns and atmospheric circulation

**FIG. 3.** The count of monthly observations used in the GPCC precipitation dataset for each 10-yr period.
related to PDSST and LTSST (Fig. 1). LTSST is related to statistically significant precipitation decreases over southwestern Asia during November–April (Fig. 5a). The significant regional precipitation decreases are observed over the Zagros Mountains, extending westward into the Middle East and Mediterranean regions, southeastern Turkmenistan, Uzbekistan, and Tajikistan. The statistically significant precipitation decreases occur over high-density precipitation reporting areas (Fig. 3). The LTSST is related to anticyclonic circulation (high pressure) extending to the north from the Indian Ocean as demonstrated by 200-hPa geopotential height (Fig. 5b). The location of this high pressure area is consistent with southwestern Asia precipitation decreases. LTSST is also related to warm surface temperatures over southwestern Asia (Fig. 5c). While the observed warming of global surface temperatures during the 1901–2004 period occur simultaneously with LTSST, the observational analysis cannot confirm to what degree LTSST contributes to surface temperature warming.

The cold phase of PDSST is related to statistically significant precipitation decreases over southwestern Asia (Fig. 5d). The statistically significant precipitation decreases occur over high-density precipitation reporting areas, notably over Tajikistan and Uzbekistan. However, the upper tropospheric atmospheric circulation related to PDSST is largely statistically insignificant and dissimilar from the circulation related to LTSST (Fig. 5e). PDSST is related to statistically insignificant high pressure over subtropical extreme East Asia and a barotropic wave train throughout Eurasia (Fig. 5d). The circulation shown in Fig. 5e are generated using the Twentieth Century Reanalysis and show considerable similarity with NCEP–NCAR Reanalysis 1 for the post-1950 period (not shown). The NCEP–NCAR Reanalysis 1 circulation indicates a stronger high pressure area over East Asia, but the barotropic wave train impinging on northern portions of southwestern Asia is pushed farther to the north. The complex regional circulation related to PDSST may displace the west-to-east storm tracks north of the region and ultimately reduce precipitation during the cold phase of PDSST over southwestern Asia. To test whether the PDSST-related changes were forced by LTSST, we removed the LTSST component of the observational data and analyzed the response to PDSST. We verified that the contributions of LTSST to the precipitation and 200-hPa geopotential height responses to PDSST were very small. PDSST is related insignificant surface temperature changes over southwestern Asia (Fig. 5f).

4. Response to steady trend forcing

Previously, we showed that LTSST is related to observed circulation modifications and precipitation decreases over southwestern Asia during November–April. Here, we examine how LTSST forces changes to the tropical Indo-Pacific precipitation and the tropospheric circulation, precipitation, and surface temperature over Asia in three atmospheric global climate models, the GFS, NSIPP-1, and GFDL AM2.1.

The atmospheric circulation responses across all three models, shown in terms of 200-hPa geopotential height (Fig. 6), to the LTSST pattern are very similar over the Middle East and southern Asia, as evidenced by their close correspondences to the three-model average (Fig. 6d). Each of the three models simulates subtropical anticyclones of the same magnitude centered at 25°N. The only intermodel difference between the anticyclonic circulations is simulated by the GFS model (Fig. 6a), which indicates that this anticyclone extends across the entirety of the Arabian Peninsula and southern Asia, instead of over just Arabia and the Arabian Sea in NSIPP-1 (Fig. 6b) and GFDL AM2.1 (Fig. 6c).

The precipitation responses (Fig. 7) to LTSST in each of the three models are also very similar over the Middle East and southwestern Asia. Each of the three models resolves precipitation decreases over southwestern Asia and the Arabian Peninsula (Fig. 7d). The modeled precipitation patterns closely resemble observed conditions over southwestern Asia (Fig. 5a) over regions in which precipitation is most heavily sampled (Fig. 3). The only intermodel differences between the precipitation responses are stimulated by the NSIPP-1 model (Fig. 7b). Despite all models resolving precipitation decreases, it is much weaker in the NSIPP-1.

Across regions bordering southwestern Asia, there is less intermodel consistency of the precipitation response
to LTSST because of the lack of strong teleconnections over those regions (Fig. 7d). In particular, over India, Bangladesh, and Myanmar, the GFS model (Fig. 7a) indicates statistically significant regional drying in response to the zonally expansive anticyclone, whereas the NSIPP-1 (Fig. 7b) and GFDL AM2.1 (Fig. 7c) indicate pluvial conditions. Over Kazakhstan and Caucasus, there are strong intermodel differences (Fig. 7d), ranging from significant drying in the GFDL model (Fig. 7c) to significant pluvial conditions in the NSIPP-1 model (Fig. 7b).

The surface temperature responses (Fig. 8) to LTSST over southwestern Asia are consistent in terms of pattern and magnitude across each of the considered models (Fig. 8d) but differ slightly across the larger domain, particularly north of the Black Sea and over India. Each of the three models simulates uniform statistically significant surface temperature increases throughout southwestern Asia and the Arabian Peninsula (Fig. 8d). The only intermodel difference between the surface temperature responses are simulated by the GFS model (Fig. 8a), which simulates surface temperature increases in excess of 1.6 K, as opposed to the range of 0.2–0.8 K simulated by the NSIPP-1 (Fig. 8b) and GFS (Fig. 8c) models. Because of the consistent pattern and magnitude of the surface temperature warming simulated by the models (Fig. 8), we conclude that a portion of the strong observed surface warming (Fig. 5c) is directly forced by LTSST.

Over India and China, the GFS model simulates very strong and statistically significant surface temperature warming, exceeding 1.6 K, in response to the strong subtropical anticyclone extending from the Arabian Peninsula to the Pacific Ocean (Fig. 8a). The NSIPP-1 model also simulates statistically significant warming over India and China, although the magnitude of
the warming is considerably less (Fig. 8b), whereas the GFDL model simulates mixed and statistically insignificant surface temperature anomalies (Fig. 8c). Over the Mediterranean and Black Sea regions, there are large intermodel differences in the surface temperature response to the SST Trend pattern (Fig. 8d). Regionally, the GFS simulates statistically insignificant warming (Fig. 8a), the NSIPP-1 simulates cooling that is statistically significant over Caucasus (Fig. 8b) and the GFDL model simulates statistically significant warming (Fig. 8c).

The tropical Indo-Pacific precipitation responses to the LTSST pattern (Fig. 9) across all three models are very similar (Fig. 9d). All three models indicate increases in precipitation over the northern Indian Ocean and Maritime Continent and decreases in precipitation over the southern Indian Ocean and tropical central Pacific Ocean. The changes in tropical precipitation occur concomitantly with changes in middle tropospheric diabatic heating, which have been shown to force anomalous atmospheric circulations over southwestern Asia on intraseasonal (e.g., Barlow et al. 2005; Hoell et al. 2013) and interannual (e.g., Barlow et al. 2002; Hoell et al. 2012) time scales. Therefore, the tropical SST and atmospheric forcing of circulations and precipitation over southwestern Asia associated with LTSST are consistent with similar forcing on intraseasonal and interannual time scales.

5. Response to steady low-frequency Pacific forcing

Previously we showed that the cold phase of PDSST is related to observed circulation modifications and precipitation decreases over southwestern Asia during November–April. Here, we examine how PDSST forces changes to tropical Indo-Pacific precipitation and the tropospheric circulation, precipitation and surface temperature over Asia in three atmospheric global climate models, the GFS, NSIPP-1, and GFDL AM2.1.

Anticyclonic circulation over Asia centered at 30°N is the predominant feature of the upper tropospheric circulation response to PDSST across all three models (Fig. 10d). However, the zonal location of the anticyclone and the meridional extent of low pressure extending from the tropics to the west of the anticyclone are simulated differently in each model (Figs. 10a–c).
The GFS model simulates the strongest anticyclone with the largest zonal extent, ranging from Iran through China into the Pacific Ocean (Fig. 10a). The NSIPP-1 (Fig. 10b) and GFDL AM2.1 (Fig. 10c) simulate anticyclonic circulation and an inverted trough to the west of the anticyclone, although the zonal location of the trough and anticyclone differ by 10° longitude.

The circulation responses to PDSST simulated by GFS, NSIPP-1, and GFDL AM2.1 share similarities (Fig. 10d), with each other and with observations (Fig. 5e) south of 40°N. The primary similarities are the anticyclone over subtropical Asia and the inverted trough feature to the west of the anticyclone. The differences in the model responses to PDSST are in part due to the PDSST forcing of the models and the method for calculating the observed response to PDSST. The PDSST EOF pattern used to force the models was very strong, scaled by plus or minus two standard deviations of the associated PC, whereas the observed response was computed through composites at plus or minus one standard deviation from the PDSST PC. The impacts of such a strong PDSST forcing pattern lead to a stronger atmospheric response in the models, but also a possible nonlinear response that could change the overall circulation.

There is general agreement among the models that PDSST forces precipitation reductions over southwestern Asia (Fig. 11), similar to observed conditions (Fig. 5d). However, there are intermodel variations in the patterns of reduced precipitation over southwestern Asia and southern portions of the plotted domain (Fig. 11d). The GFS (Fig. 11a) and GFDL models simulate drying over southwestern Asia with particularly severity and statistical significance over Pakistan, Afghanistan, and Tajikistan. Over western and southern portions of southwestern Asia (viz. Iraq and Saudi Arabia), the GFS and GFDL models simulate drying on the large scale with some subregional variations. The NSIPP-1 (Fig. 11b) model simulates precipitation increases over southern Asia due to a large scale inverted trough extending from the Indian Ocean into southern Asia (Fig. 10b). However, the NSIPP model simulates drying in response to PDSST over northern portions of southwestern Asia (Fig. 11b), outside of the influence of the inverted trough (Fig. 10b).

There is very little agreement between the surface temperature responses to PDSST simulated by each model, as evidenced by the three-model average shown in Fig. 12d. The GFS model (Fig. 12a) simulates...
statistically significant warming over southwestern Asia and Asia north of 25ºN due to the position of strong high pressure throughout the region (Fig. 10a). On the contrary, the GFDL model (Fig. 12c) simulates significant cooling over southwestern Asia as a result of the inverted trough extending from the tropics into southern Asia. Furthermore, the NSIPP-1 model (Fig. 12b) simulates a statistically insignificant surface temperature
response over southwestern Asia, presumably because of the weak inverted trough simulated by the model (Fig. 10b) just to the south of the region. Overall, the model consensus of a weak surface temperature response to PDSST (Fig. 12d) is similar to observation (Fig. 5f).

The tropical Indo-Pacific precipitation responses to the PDSST pattern across all three models are very similar (Fig. 13d). All three models indicate increases in precipitation over the Maritime Continent and decreases in precipitation over the central tropical Indian Ocean and central tropical Pacific Ocean. The pattern of tropical Indo-Pacific precipitation forced by LTSST is very similar to the precipitation patterns known to force southwestern Asia climate on intraseasonal to interannual time scales.

The tropical Indo-Pacific precipitation response patterns to PDSST (Fig. 13d) and LTSST (Fig. 9) are similar. However, the magnitudes of the precipitation forced by LTSST and PDSST are considerably different, as the PDSST pattern forces much stronger Indo-Pacific precipitation changes. The primary reason for the increase in tropical precipitation forcing by PDSST relative to LTSST is in the magnitude of the forcing generated by the DWG. The scaling factor of the SSTs in the PDSST was plus or minus two standard deviations while the scaling factor of the SSTs in the LTSST experiment was plus or minus one standard deviation. Because of the differences in the magnitude of the tropical Indo-Pacific precipitation, and therefore diabatic forcing, there may be nonlinear differences in the atmospheric responses to LTSST and PDSST. These nonlinear differences are likely one reason for the different teleconnections over Asia forced by LTSST and PDSST.

6. Summary and discussion

The complex societal vulnerabilities of southwestern Asia, an area containing the nations of Iraq, Iran, Afghanistan, and Pakistan, are exacerbated by severe droughts during the primary precipitation season of November–April. November–April precipitation over southwestern Asia, accounting for roughly 75% of the annual total, has demonstrated decadal and multidecadal variability during the twentieth century (Fig. 4). In light of southwestern Asia’s sensitivity to Indo-Pacific SST variability on interannual time scales, we examine the effects of the long-term trend in global SST (LTSST) and pan-Pacific decadal SST (PDSST) variability on the regional circulation and precipitation during November–April.

Observed southwestern Asia precipitation is poorly sampled in space and time (Figs. 2b and 3). Even during
time periods in which southwestern Asia precipitation was most frequently sampled, precipitation observations were taken in only a few locations, namely, the periphery of Iran, portions of Turkmenistan and Uzbekistan, and over Tajikistan. Because of the spatial coverage of precipitation sampling, there are large uncertainties in the regional precipitation variability, particularly over Iraq, Afghanistan, and Pakistan. Despite the poor regional coverage of precipitation observations, Schiemann et al. (2008) reported that gridded datasets such as GPCC used in this analysis have the capacity to qualitatively identify wet and dry years.

The DWG identified the SST patterns characteristic of LTSST (Figs. 1a,b) and PDSST (Figs. 1c,d) using empirical orthogonal functions (Schubert et al. 2009). The LTSST and PDSST patterns were used to force a suite of global climate models (Schubert et al. 2009) to better understand the atmospheric response to the two primary modes of decadal to multidecadal global SST variability. Additionally, we performed observational analyses to compare with the suite of model experiments.

Observational analysis (Figs. 5a–c) and models forced by LTSST during November–April indicate that LTSST is associated with anticyclonic circulation over southwestern Asia (Fig. 6) that results in statistically significant regional precipitation decreases (Fig. 7) and surface temperature increases (Fig. 8). For PDSST, observational and modeling analyses marginally agree on the atmospheric circulation response (Figs. 5d–f). PDSST forces small variations of particular features, specifically the location of anticyclonic circulation over Asia and an inverted trough to the west of the anticyclone. These small changes can have large consequences on precipitation over southwestern Asia.

November–April snowfall in the high mountains of southwestern Asia melts during the springtime and plays a critical role in river flows and regional vegetation. On interannual time scales, ENSO has been linked to hydroclimatic variability over southwestern Asia, specifically in river flow (Barlow and Tippett 2008), soil moisture (Sheffield and Wood 2008), and the normalized difference vegetation index (Kariyeva and van Leeuwen 2012). Considering the similarities between the atmospheric circulations and precipitation responses between PDSST, LTSST, and those associated with ENSO, it is reasonable to deduce that low-frequency Pacific variability has strong and potentially similar effects on the regional hydroclimatic variability over southwestern Asia, but over longer time scales. Therefore, an improved understanding of the land surface hydrology related to decadal and multidecadal climate variability over southwestern Asia is of critical importance.

![Fig. 11. As in Fig. 7, but for between Pdc and PDw.](image-url)
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