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Atlantic effects on recent decadal trends in global monsoon

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Abstract

Natural climate variability contributes to recent decadal climate trends. Specifically the trends during the satellite era since 1979 include Atlantic and Indian Ocean warming and Pacific cooling associated with phase shifts of the Atlantic Multidecadal Oscillation and the Pacific Decadal Oscillation, and enhanced global monsoon (GM) circulation and rainfall especially in the Northern Hemisphere. Here we evaluate effects of the oceanic changes on the global and regional monsoon trends by partial ocean temperature restoring experiments in a coupled atmosphere–ocean general circulation model. Via trans-basin atmosphere–ocean teleconnections, the Atlantic warming drives a global pattern of sea surface temperature change that resembles observations, giving rise to the enhanced GM. The tropical Atlantic warming and the resultant Indian Ocean warming favor subtropical deep-tropospheric warming in both hemispheres, resulting in the enhanced monsoon circulations and precipitation over North America, South America and North Africa. The extratropical North Atlantic warming makes an additional contribution to the monsoon enhancement via Eurasian continent warming and resultant land–sea thermal gradient over Asia. The results of this study suggest that the Atlantic multidecadal variability can explain a substantial part of global climate variability including the recent decadal trends of GM.

Keywords: Global monsoon, PDO, AMO, meridional thermal gradient

1. Introduction

Monsoons develop over tropical-to-subtropical land and adjacent ocean regions (North and South America, North and South Africa, South and East Asia and Australia) and are characterized by salient seasonal variations in atmospheric circulation and precipitation (e.g. Murakami and Matsumoto 1994; Webster et al. 1998; Webster 2006). These seasonally reversing atmospheric overturning circulations throughout the tropics may be viewed as a "global monsoon" (GM) system (e.g. Trenberth et al. 2000; Wang and Ding 2008; Wang et al. 2014; An et al. 2015). The GM system varies on a wide range of timescales both naturally and in response to external forcing such as Earth's orbital parameters (associated with paleo-monsoon variability; Wang et al. 2014; Jiang et al. 2015; Yan et al. 2016) and anthropogenic radiative forcing agents (detailed below). Christensen et al. (2014) reviewed historical variations and future projections of GM by atmosphere–ocean coupled general circulation models (AOGCMs; e.g. Hsu et al. 2012, 2013; Kitoh et al. 2013; Lee and Wang 2014; Endo and Kitoh 2014). Since the early 20th century, substantial decadal-to-multidecadal variations in the GM subsystems have been reported, implying the importance of natural climate variability for historical GM variations.

The large-scale horizontal thermal gradient between land and ocean is a primal driver of the monsoon system. Meridional thermal gradient (MTG), measured by the gradient of middle to upper tropospheric temperature over South Asia and the tropical Indian Ocean (IO), is an useful index for Asian monsoon intensity (Li and Yanai 1996; Kawamura 1998; Arai and Kimoto 2008; Dai et al. 2013). Monsoon itself also affects the MTG via condensation heat release in the troposphere, resulting in a positive feedback between the monsoon and MTG (e.g. Li and Yanai 1996). Such deep-tropospheric land–sea thermal gradient tracks Asian monsoon variability (Zuo et al. 2013; Roxy et al. 2015), projected future change (Ueda et al. 2006; Sun et al. 2010; Dai et al. 2013; Ma and Yu 2014; Ogata et al. 2014; Kamae et al. 2014b; Sooraj et al. 2015), and past climate changes (e.g. Ueda et al. 2011; Man et al. 2012).

The deep-tropospheric thermal gradient and monsoon systems are sensitive to spatial patterns of aerosols forcing and sea surface temperature (SST) anomaly (e.g. Kamae et al. 2015; Dong et al. 2016). Anthropogenic aerosols affect regional precipitation (e.g. Rotstayn and Lohmann 2002; Hwang et al. 2013; Wang et al. 2016) over the Asian (e.g. Bollasina et al. 2011; Dong et al. 2016) and African monsoon regions (e.g. Rotstayn et al. 2000; Rotstayn and Lohmann 2002; Kawase et al. 2011). Polson et al. (2014) pointed out that aerosols are fundamental drivers for the observed long-term weakening trend in GM rainfall since the mid-20th century. They also noted that the current AOGCMs still have limited representation of indirect effects of aerosols (Stevens and Feingold 2009), indicating that total aerosols effects on monsoons are still uncertain. Decadal trends in anthropogenic aerosols are weak since 1979 (Fig. 1b in Polson et al. 2014), a period with increased GM rainfall (Wang and Ding 2006; Hsu et al. 2011; Wang et al. 2012, 2013; Polson et al. 2014) particularly over the Northern Hemisphere (NH).

The recent decadal SST trend since 1979 (Fig. 1a) are influenced by basin-scale multidecadal variability modes: Interdecadal Pacific Oscillation (IPO) or Pacific Decadal Oscillation (PDO) and Atlantic Multidecadal Oscillation (AMO; e.g. Power et al. 1999; Deser et al. 2004; Newman et al. 2016). The tropical SST trends feature warming over the tropical Atlantic, IO, and western Pacific (IOWP) and cooling over the eastern Pacific (e.g. Luo et al. 2012; Watanabe et al. 2014; McGregor et al. 2014; Li et al. 2016). These SST trends are of great importance for the decadal trends in global-mean surface air temperature (e.g. Kosaka and Xie 2013; Watanabe et al. 2013; Li et al. 2016) and global atmospheric circulations including monsoons (Wang et al. 2012, 2013; Trenberth et al. 2014; Ueda et al. 2015). Through partial ocean temperature restoring experiments, McGregor et al. (2014), Li et al. (2016), and Kucharski et al. (2016) showed that tropical Atlantic SST can explain a substantial part of the recent decadal variations of the tropical SST and atmospheric overturning circulation. The partial temperature restoring technique is effective to examine local and remote influences of the oceans on the monsoons and avoids the ambiguity in causality of SST variations over the warm oceans that is present in atmosphere-only simulations with prescribed SST (Wang et al. 2005).

This study examines how recent SST change (for 1979–2012) in each tropical ocean basin (Atlantic, IO and Pacific) affects the GM and its subsystems through the ocean temperature restoring experiments. Atmospheric general circulation models are often used to quantify SST contributions from individual basins but the results do not consider coupled inter-basin interactions; on interannual timescales, for example, SST anomalies over the IO and Atlantic are forced by El Niño Southern Oscillation (ENSO) in the Pacific (e.g. Chikamoto et al. 2015; Li et al. 2016). In addition, the IO and Atlantic SST also affect the ENSO amplitude and frequency (Terray et al. 2015). Here we adopt the partial coupling method to explore the effect of cross-basin interactions on the recent trends of the GM. We also apply the MTG-view for a physical interpretation on the decadal GM trend. Section 2 describes the data and methods including model experiments and observations used in this study. Section 3 presents the decadal climate trends over the tropics in SST and monsoon precipitation.

Section 4 compares regional patterns of trends between MTG, monsoon circulation and precipitation. Section 5 is a summary with discussions.

2. Data and methods

2.1. Observations and reanalysis

To examine decadal climate trends for 1979–2012, we used observed SST dataset (HadISST; Rayner et al. 2003) which was also used for model simulations (Sect. 2.2). We used three precipitation datasets (CRU TS v3.23; Harris et al. 2014, Global Precipitation Climatology Project; hereafter GPCP; Huffman et al. 2009, and the CPC Merged Analysis of Precipitation; hereafter CMAP; Xie and Arkin 1997) to evaluate observational uncertainty (i.e. spread among datasets) in decadal trend of summertime precipitation. Three dimensional atmospheric variables were derived from ERA-Interim (Dee et al. 2011) that well reproduces the interannual variability and trend of the GM system since 1979 (Lin et al. 2014).

The monsoon domains are determined by an annual range of precipitation. In this study, areas where the annual range of climatological GPCP precipitation (1979–2012) is greater than 2.5 mm day^{-1} are defined as monsoon domains, similar to Kitoh et al. (2013). We mainly focus on land monsoon to examine remote influence of restored ocean temperature on land rainfall (Sect. 2.2). The GM domain consists of seven subdomains: North American Monsoon (NAM); South American Monsoon (SAM); North African Monsoon (NAF); South African Monsoon (SAF); South Asian Monsoon (SAS); East Asian Monsoon (EAS); and Australian Monsoon (AUS; see

Sect. 3.2). For regional division, 20°N and 100°E are used to separate SAS from EAS (Fig. 5 in Kitoh et al. 2013).

In this study, decadal trends are calculated by low-pass-filtered (3-yr running mean) variables for 1979–2012. The 3-yr running mean can largely remove interannual variability including the effect of ENSO (Wang et al. 2013) and increase signal-to-noise ratio in decadal trends. We confirmed that large-scale patterns of decadal trends were not sensitive to smoothing time windows (e.g. 5-yr running mean) of low-pass filter. We examine summer monsoons by local summer mean determined by May to September in the NH and November to March in the Southern Hemisphere (SH).

2.2. Model simulations

The National Center for Atmospheric Research (NCAR) coupled climate model, the Community Earth System Model version 1.06 (CESM1.06; Hurrell et al. 2013) was used to investigate global influences of the observed basin-scale ocean temperature trends for 1979–2012. This study focuses on influences of five basin-scale ocean temperature: Atlantic (Atl run), IO (IO run), Pacific (Pac run), tropical Atlantic (tAtl run) and tropical Pacific (tPac run; detailed below). The atmospheric component of this model is the Community Atmospheric Model version 4 (CAM4; Neale et al. 2013) with F19_G16 horizontal resolution (~2°). Oceanic component has ~1° horizontal resolution. The basic experimental framework of this work is generally similar to Li et al. (2016). We restored basin-scale ocean mixed-layer temperature in the coupled model as follows:

$$F = cD\frac{T_r - T_m}{\tau}, \tag{1}$$

where c is the heat content of sea water, D is the mixed-layer depth, T_r is the restoring target temperature, T_m is the model temperature, and τ is the restoring timescale (10 days). We added external heating F to the model to restore basin-scale ocean mixed-layer temperature.

The climate response to the restored ocean mixed-layer temperature was calculated by the difference between a control run and a perturbed run. In the control run, the mixed-layer temperature was restored to the model climatology. In the perturbed run, observed temperature trend for 1979–2012 was added to the mixed-layer temperature restored in the control run. We conducted 12-member ensemble simulations with slightly different initial conditions in the control and perturbed runs. The model was integrated for 15 years and the last 10 years were used for analyses. The Atlantic, Pacific, and IO temperature were restored for the Atl, Pac and IO runs. In addition to the whole ocean basin restoring runs, we also conducted tropical-only restoring runs for the Atlantic (tAtl run) and Pacific (tPac run). In the tropical runs, the mixed-layer temperature were restored in the tropical region (20°S–20°N). Strength of temperature restoring was gradually reduced in subtropical buffer regions (30°S–20°S and 20°N–30°N).

2.3. CMIP5 multiple model ensemble

We also used the Coupled Model Intercomparison Project phase 5 (CMIP5) multiple model dataset (Taylor et al. 2012) for comparison with the CESM simulations and observations. Model data from the historical and representative concentration pathway (RCP) 4.5 runs (Meinshausen et al. 2011) were used for 1979–2005 and 2006–2012 periods, respectively. We used the results obtained from 24 models listed in Table S1 in the online supplement. All the CMIP5 model outputs were interpolated into a 2.5° × 2.5° horizontal grid before analyses.

3. Decadal trends in summertime SST and monsoon rainfall

3.1. SST

We first examine decadal SST trends simulated in the ocean temperature restoring experiments. Physical processes associated with simulated SST due to the restored tropical ocean temperature were detailed in Li et al. (2016). Figure 1a shows observed summertime (May to September in the NH and November to March in the SH) SST trend for 1979–2012. As suggested in previous studies (e.g. Kosaka and Xie 2013; Li et al. 2016), the tropical zonally asymmetric SST trend (warming over the tropical Atlantic and IOWP and cooling over the eastern Pacific associated with positive AMO and negative PDO or IPO) is apparent. In contrast to observations, CMIP5 multi-model mean shows an overall warming without any regional cooling in the tropics (Fig. 1e), indicating the importance of natural climate variability to the observed trends (e.g. Kamae et al. 2014a, 2015). Figures 1b–d and 2 show results of the basin-scale ocean temperature restoring runs. When the North Atlantic warming is restored, the model can reproduce the general characteristics of the SST trends (i.e. the tropical zonally asymmetric SST trend; Figs. 1b, 2a; Li et al. 2016). The Gill-type atmospheric response to the tropical Atlantic warming results in surface easterly and westerly anomalies over the IO and the eastern Pacific (Fig. S5 in Li et al. 2016), leading to sea surface warming and cooling through the wind–evaporation–SST feedback (Xie and Philander 1994). In addition, the Bjerknes feedback amplifies the east–west ocean temperature asymmetry (the western Pacific warming and the eastern Pacific cooling; Figs. 1b, 2a; Li et al. 2016). Restoring extratropical Atlantic temperature (difference between the Atl and tAtl runs) results in additional sea surface warming over the tropical western Pacific (150°E–180°; Fig. 1b) that is similar to the observations. Note that the Atl and tAtl runs do not reproduce accurately several parts of the observed SST trends (e.g. simulated cooling in the eastern tropical Pacific is weaker than the observations). Spatial coherence between the observed and simulated SST trends in the IO restoring run is limited although the western Pacific and the extratropical Atlantic warming are reproduced in this run (Fig. 1c). The Pac run cannot reproduce SST trends over the tropical Atlantic and the IO (Figs. 1d, 2b), at least in the ensemble mean. Physical mechanisms responsible for the global SST response and atmospheric circulation to the ocean temperature restoring were detailed in Li et al. (2016; see also Luo et al. 2012; Kosaka and Xie 2013; McGregor et al. 2014). Next we examine decadal trends in precipitation found in the observations and model simulations.

3.2. Global monsoon rainfall

Figures 3–5 show observed and modeled decadal precipitation trends during local summer. The observed trend since 1979 (Fig. 3a–c) exhibits regional patterns (detailed below) associated with the observed global SST trend (Fig. 1a), the AMO and PDO (e.g. Trenberth et al. 2014; Gu et al. 2016). Generally the land monsoon rainfall shows an increasing trend over the NH (Wang et al. 2013) and SH (Fig. 5a). Note that different observation datasets exhibit a large spread (Fig. 5a) particularly over the oceans (Fig. 3a–c; Wang et al. 2013, Sperber et al. 2013 Hao et al. 2016), suggesting a non-negligible observational uncertainty. Table 1 summarizes correlation coefficients of the tropical (30°S–30°N) precipitation trends between the observations and the model simulations. The model experiments cannot reproduce accurately several parts of observed regional precipitation trends (detailed in Sects. 3.3 and 3.4), resulting in limited correlation coefficients. Generally correlations in the Atl (0.26 and 0.30 with GPCP and CMAP, respectively) and Pac (0.33 and 0.15 with GPCP and CMAP, respectively) runs are higher than the IO run and CMIP5 multi model mean (detailed below). Although the ocean temperature restoring in the Atl run is limited to the narrow area (Fig. 1), substantial remote influence on the other ocean basins and tropical land results in the high spatial correspondence of the precipitation trend. In this study, we mainly focus on land monsoon rainfall, which shows increasing trends consistently among different observational datasets except for NAF (decrease in CMAP), SAS and SAM (decrease in GPCP; Fig. 5a).

The Atl/tAtl runs can reproduce the increasing trends of NH, SH monsoons and GM rainfall (Fig. 5). The restored Atlantic temperature in these simulations contributes to the general increase in land rainfall and its spatial patterns except SAM, SAS, EAS and AUS domains (Figs. 3–5). In contrast to the Atl/tAtl runs, IO and Pac/tPac runs do not reproduce the increasing trends in GM, NH and SH rainfall (Fig. 5b). However, restored IO and Pacific Ocean temperature also lead to positive or negative trends of the regional monsoon rainfall (Fig. 5b), suggesting their partial contributions to the regional rainfall trends (detailed in Sects. 3.3 and 3.4). Here the forced response derived from the CMIP5 multi model mean shows a limited precipitation trend globally and regionally (Fig. 3g and Fig. S1 in the online supplement). Spatial correlation between the CMIP5 multi model mean and the observations is small (Table 1), indicating the importance of internal climate variability including the AMO and PDO. In the later subsections, we further examine regional trends of monsoon rainfall and associated physical processes.

3.3. African and American monsoon rainfall

Over tropical North Africa, the observations consistently show decreasing and increasing trends in precipitation over the equator and the tropical monsoon domain (off the equator, ~10°N), respectively (Fig. 3a–c). This northward shift of the precipitation band results in the increased NAF rainfall except in CMAP (Fig. 5a). This increased rainfall is reproduced in the Atl/tAtl runs (Figs. 3d, 4a, 5b) but not in the IO and Pac runs.

The recent positive AMO-related warm tropical North Atlantic (Fig. 1) affects the position of the ITCZ and strength of the NAF rainfall (Folland et al. 1986; Zhang and Delworth 2006; Ting et al. 2011; Rodríguez-Fonseca et al. 2015; Kamae et al. 2016b). The IO warming in the IO run (Fig. 1) causes a reduction of NAF rainfall (Figs. 3e, 5b), consistent with the suppression of tropical African rainfall by westward propagating Rossby waves (Lu 2009) initiated by the IO warming (Figs. 3e, 5b; e.g. Hoerling et al. 2006). The observed trends in the Pacific sea water temperature restored in the Pac/tPac runs result in reductions of NAF and SAF rainfall (Figs. 3f, 4b, 5b). While previous studies emphasized the effect of the observed eastern Pacific cooling on Sahel rainfall (e.g. Hoerling et al. 2006), the decreased Sahel rainfall is also affected by tropical SST anomalies elsewhere (e.g. warm tropical IO). Mohino et al. (2011) showed that the tropical African rainfall anomaly is quite limited if only Pacific SST anomalies are included without SST anomalies in other ocean basins (Figs. 10–11 in Mohino et al. 2011). Our results suggest that the observed Pacific Ocean temperature does not substantially affect SST anomalies of the tropical IO and Atlantic (Figs. 1, 2) and has a limited influence on tropical African rainfall.

The observations show the largest increasing trends in NAM rainfall among the NH monsoons (Fig. 5a). Only the Atl/tAtl runs reproduce this increasing trend quantitatively (Fig. 5b). The larger North Atlantic warming compared with the South Atlantic (Fig. 1) results in an enhanced Caribbean precipitation (Figs. 3d, 4a; Ting et al. 2011). During the austral summer, the Atlantic warming results in an increase in near-equatorial South American rainfall (contributing to an increase in SAM rainfall) and a decrease in subtropical (~20°S) South American rainfall (Figs. 3d, 4a, 5b), consistent with Kushnir et al. (2010). The large and significant rainfall increases in NAM, NAF and SAM in the Atl/tAtl runs contribute to the increasing trends in the NH, SH and GM rainfall (Fig. 5b), consistent quantitatively with the observations (0.19–0.50 mm day^{-1} 34 yr^{-1} in the observations and 0.18 ± 0.02, 0.19 ± 0.03 mm day^{-1} 34 yr^{-1} in the Atl and tAtl runs, respectively). In contrast, the IO run shows reductions of monsoon rainfall over several domains (NAF, NAM, SAM and SAS; detailed in Sect. 3.4), resulting in a suppression of GM rainfall (Fig. 5b). These negative trends suggest that the IO warming compensates partly these observed positive trends. The Pacific Ocean temperature also compensates the positive trends of NAM, NAF and SAF rainfall and amplifies the positive SAM rainfall trend (Fig. 5b).

3.4. Asian and Australian monsoon rainfall

The precipitation trend over the Asian monsoon regions exhibits a complex spatial pattern (Fig. 3a–c). The warming trend over the IOWP (Fig. 1a; Solomon and Newman 2012; Roxy et al. 2014) induces a local increase in precipitation over the ocean (Fig. 3a, b; Zhou et al. 2009; Wang et al. 2012; Ueda et al. 2015). Summertime precipitation is suppressed over the central-east and northern regions of the Indian subcontinent and is increased over the southwestern regions (Figs. 3a–d, 4a) associated with the tropical (the Atlantic and IOWP) oceanic warming (Figs. 1a, b, 2a; Mishra et al. 2012; Roxy et al. 2015). These regional variations in precipitation trends are reproduced in the Atl/tAtl runs (Figs. 3d, 4a). Over the SAS and EAS domains, the observed land monsoon rainfall shows generally positive trends (not statistically significant in GPCP; Figs. 3a, 5a). The Atl/tAtl runs do not show systematic trends over the SAS and EAS domains (Figs. 3d, 5b), suggesting a limited contribution of the Atlantic Ocean to the rainfall trends over the Asian land monsoon domains.

Atlantic temperature influences on Indian monsoon rainfall through the shift of ITCZ, Eurasian continental warming and mid-latitude wavetrain (Goswami et al. 2006; Li et al. 2008; Wang Y et al. 2009; Ting et al. 2011). Zhang and Delworth (2006) reported a relationship between the North Atlantic warming and the increased Indian rainfall, which is not apparent in our simulations (Figs. 3d, 4a). The Atlantic warming also results in an IO warming (Fig. 1b) through eastward propagating Kelvin wave (Kucharski et al. 2009; Wang C et al. 2009; Li et al. 2016). Here the IO warming restored in the IO run and Eurasian warming produces the opposite effects on the Asian monsoon via changing MTG (Zuo et al. 2013). The IO warming also leads to an enhanced water vapor advection that can increase South Asian monsoon rainfall (Roxy et al. 2015; Ueda et al. 2015). Section 4 further examines the competition between the IO and Eurasian continental warming in the MTG and the Asian monsoon.

The Pac/tPac runs with eastern-cool/western-warm SST anomaly over the tropical Pacific simulate the increased western North Pacific and SAS rainfall (Figs. 3f, 4b, 5). The relationship between the Pacific zonally asymmetric SST anomaly and the Indian rainfall is consistent with previous reports (Rasmusson and Carpenter 1983; Webster and Yang 1992; Kawamura 1998; Kumar et al. 1999; Mishra et al. 2012). However, the contributions of the Pac/tPac to NAM, NAF, SAM, and SAF rainfall are limited (Figs. 3f, 4b, 5b), resulting in an inconsistent trend of GM rainfall with the observations (Fig. 5). The IO warming restored in the IO run causes to increase the local rainfall (Fig. 3e; Xie et al. 2009; Ueda et al. 2015) but suppresses rainfall over the interior of the Indian subcontinent via reducing MTG between the Asian continent and the tropical ocean (Fig. 3e; see Sect. 4; Mishra et al. 2012; Roxy et al. 2015). In short, the general reduction of land monsoon rainfall simulated in the IO run (Figs. 3e, 5b) can largely be explained by the suppression of interior continental rainfall over India, reduction of adjacent tropical African rainfall (Hoerling et al. 2006), and reduced NAM and SAM rainfall. The increase in SAS rainfall in the Pac run is larger than the Atl run (Fig. 5b), consistent with differences in the simulated SST between the two runs (the weakened eastern Pacific cooling and warmed IO in the Atl run; Fig. 1).

All the observation datasets exhibit substantial increasing trend in the AUS rainfall (1.2–1.4 mm day^{-1} 34 yr^{-1}; Fig. 5a). None of the experiments reproduce the AUS rainfall increase (Figs. 3, 4, 5b) despite strong controls of the IO and Pacific SST anomalies on northern Australian rainfall (e.g. Cai and van Rensch 2013). Rotstayn et al. (2007) pointed out that aerosols are essential factors for the observed increasing trend of summertime rainfall over northwestern Australia (Smith 2004; Taschetto and England 2009) since the late 20th century. The aerosols effect is not explicitly included in our simulation, where aerosols are held constant in time.

4. Meridional thermal gradient and global monsoon

Variations in the horizontal temperature gradient are suggested to be important for interpretation of large-scale variations in the monsoon strength on different timescales: seasonal to interannual (Li and Yanai 1996; Kawamura 1998), decadal (Zuo et al. 2013), and centennial to millennium (Ueda et al. 2006, 2011; Man et al. 2012; Zuo et al. 2013; Roxy et al. 2015; Mohtadi et al. 2016). In this section, we advance our interpretation on dynamic and thermodynamic properties of the recent decadal GM trend. Figure 6a shows climatologies of mid-to-upper tropospheric thickness (between 200 and 500 hPa levels) and vertical wind shear between lower and upper troposphere (850 hPa minus 200 hPa). Horizontal gradient of climatological thickness corresponds to deep tropospheric temperature gradient. Through a thermal wind balance, variation in the thickness corresponds well with an index for seasonal cycle and interannual variability of the South Asian monsoon (Webster-Yang index) determined by the tropospheric wind shear (Webster and Yang 1992). The tropospheric thickness and vertical wind shear can explain changes in atmospheric thermodynamic structure and atmospheric stability in a perturbed climate (Ma et al. 2012; Kamae et al. 2016a). Thick lines in Fig. 6a indicate latitudes with peak thickness in each hemisphere. Peak latitudes over longitudinal areas with continents (and associated monsoon systems) are displaced from the equator. Here monsoonal circulations are formed along with resultant meridional peak-to-equator thickness gradient (i.e. MTG). Over South Asia, the peak latitude is farthest from the equator, the peak thickness (i.e. the South Asian High) is largest and the Asian monsoon is strongest over the globe (e.g. Li and Yanai 1996). The peak latitude is located on the equator over longitudinal areas without any monsoon systems (e.g. the central Pacific).

Figure 6b shows trends of the thickness and wind shear for 1979–2012 derived from the reanalysis. A recent trend in the mid-to-upper tropospheric thickness can be characterized by increases over subtropical continents (Fig. 6b) and decreases over the tropical central-to-eastern Pacific associated with the tropical Pacific cooling (Kamae et al. 2015). Vertical wind shear (low-level westerly and upper-level easterly) is strengthened over the tropical NH (0°N–15°N; Wang et al. 2013), corresponding to the positive MTG trend (Fig. 6b). We can examine trends of the monsoon strength by comparing the thickness trends on the climatological peak latitude (dashed line in Fig. 6b) and the equator (detailed below). Most parts of the subtropics except the western Pacific exhibit larger positive thickness trend than the equator, indicating enhanced MTG. This positive MTG trend can explain well the trend in wind shear (Fig. 6b). Here the increasing trends in MTG and wind shear are consistent with the positive trend in the monsoon rainfall (Fig. 5).

Figures 6c–e and 7 show results of the individual ocean temperature restoring runs. The Atl/tAtl runs reproduce the GM rainfall increase with the subtropical peaks of thickness trends over the North and South Atlantic and South IO. Figure 8 shows MTG trends as represented by the difference in thickness trends over the peak latitude (black line in Fig. 6) and the equator. The observed MTG trends are positive except 100°E–160°E in the NH and except 140°E–180°E in the SH, generally consistent with the positive trend in GM rainfall (Fig. 5). The Atl/tAtl runs reproduce warming trends over the tropical Atlantic and IOWP (Figs. 1b, 2a) and associated regional tropospheric warming (Figs. 6c, 7a). For example, the recent tropical Atlantic warming favors an eastward propagating Kelvin wave and subtropical westward propagating Rossby waves (Kamae et al. 2014a; Li et al. 2016), resulting in (1) the positive thickness trends over subtropical North and South America, North and South Atlantic, and North Africa; and (2) the positive thickness trends over the tropical IO (Figs. 6c, 7a) associated with the IO warming (Figs. 1b, 2a) through a wind–evaporation–SST feedback (i.e. reduced scalar wind speed over the tropical IO results in sea surface warming; Xie and Philander 1994; Li et al. 2016). The Atl/tAtl runs reproduce the NH and SH subtropical peaks of the thickness trends over the Atlantic (Figs. 6c, 7a) and the positive MTG trend from the Western Hemisphere to 50°E (Fig. 8a). Note that the simulated trends over the SH are not statistically significant and are smaller than the observations (Fig. 8b).

The tAtl run simulates the recent trends in the MTG and the wind shear, similar to the Atl run, indicating an importance of the tropical Atlantic temperature trend in the dynamic and thermodynamic aspects of the recent GM enhancement (Figs. 7a, 8). The extratropical North Atlantic warming is absent in the tAtl run (Fig. 2a), leading to differential tropospheric warming over the NH extratropics including North Africa and Eurasian continent (Fig. 7a; e.g. Lu et al. 2006; Wang Y et al. 2009; Zuo et al. 2013) and resultant MTG trend over 10°W–140°E (Fig. 8a; negative and positive in the tAtl and Atl runs, respectively). This result is qualitatively consistent with the difference in precipitation trends over NAF and SAS domains between the two runs (larger in the Atl run than the tAtl run; Fig. 5b).

The Atl/tAtl runs exhibit different MTG trends to the observations over the SAS domain (Figs. 6, 8a). The observed thickness trend over West, South and East Asia exhibit a wavy pattern (Fig. 6b), suggesting an importance of mid-latitude atmospheric internal variability (e.g. Kamae et al. 2014a, 2016b). In the Atl run, the Eurasian warming is underestimated (Fig. 6c) and associated MTG trend is opposite to the observations (Fig. 8a) although SST warming over the North Atlantic and tropical IO are reproduced well (Fig. 1). Similarly, the Atl run exhibits an opposite trend in atmospheric circulation (easterly shear; Fig. 6c) to the observation (westerly shear; Fig. 6b) over the tropical IOWP (EQ–15°N; 50°E–130°E). These results are consistent with different land precipitation trends over South Asia between the Atl run and the observations (Figs. 3d, 5b). In contrast, the Pac/tPac runs exhibit larger thickness trend over the Eurasian continent than over the tropical IO, intensifying the Asian monsoon circulation and SAS rainfall (Figs. 3f, 4b, 5b, 6e, 8b). These results indicate that the trans-basin teleconnection initiated by the Atlantic Ocean contributes to the recent intensification of GM but cannot explain some of the observed trends including those over the SAS. In the SH, the Atlantic temperature also contributes to the enhanced SAM through the tropical warming (Figs. 5b, 6c, 7a, 8) but cannot explain generally the recent AUS rainfall increase, to which aerosols might be important (Rotstayn et al. 2007; see Sect. 3.4).

5. Summary and discussion

We systematically examine contributions of large-scale decadal climate variability to the recent enhancement of GM from a dynamic-thermodynamic perspective. The North Atlantic warming contributes to the regional patterns of SST trend over the tropics and associated increasing trend of GM rainfall through the trans-basin atmosphere–ocean interactions. The warm tropical Atlantic reinforces the monsoon circulations and rainfall over NAM, SAM and NAF domains. In addition, the extratropical North Atlantic warming contributes to the increasing trend of MTG over South Asia via Eurasian continental warming, but it cannot explain the increased SAS rainfall quantitatively. The increased AUS rainfall cannot be reproduced by any of the ocean temperature restoring experiments conducted in the current study, supporting the hypothesis that anthropogenic aerosols are essential for the recent increasing trend in AUS rainfall. The great importance of the Atlantic-induced pan-tropical decadal variability in the GM system suggests a possible weakening of the GM in accordance with a phase shift of the Atlantic multidecadal variability mode in the near future.

The results of this study indicate the great contribution of the Atlantic warming to the several aspects of the recent decadal climate trends including monsoons. However, the Atlantic temperature restoring experiment performed in this study cannot reproduce parts of the observed SST trends. For example, SST trends over the eastern tropical and subtropical Pacific (cooling over the tropics and warming over the middle latitude) are generally underestimated (Fig. 1a, b). Such regional differences in SST trends compared with the observations possibly affect land monsoon rainfall via large-scale atmospheric teleconnections (e.g. Pacific-North American pattern; Mo and Livezey 1986) and atmospheric thermodynamic structure (Fig. 6). We should also keep in mind that the AOGCM is not necessarily perfect in simulating trends of atmospheric circulation and precipitation even if the AOGCM can simulate accurately the global SST trends. Additional experiments including global ocean temperature restoring experiment may be effective to separate effects of the regional SST biases, model's own biases, and the missing external forcing agents (i.e. aerosols). It is also needed to compare the results of this study with similar restoring experiments performed by other AOGCMs to address robustness of the results.

Future modeling efforts by multiple climate models may improve our understanding of importance of the trans-basin atmosphere–ocean interactions in the trends of global climate system including monsoons.

Via restoring ocean temperature, Wang et al. (2013) also pointed out the importance of the North Atlantic warming in the recent intensification of the NH monsoons. The current study provides a more comprehensive view on the individual basin-scale ocean temperature contributions to the GM trend. The use of MTG can offers a physical interpretation of decadal variabilities of the global and regional monsoons. The recent regional MTG trends correspond well to the summertime precipitation trends for the last 34 years. We should note that MTG is not necessarily a universal index for long-term rainfall change due to an importance of thermodynamic rainfall change in a warming climate despite the robust dynamic rainfall change (e.g. Ueda et al. 2006; Endo and Kitoh 2014). Further research into monsoon variability and changes helps advance predictive understanding of changes in global atmospheric circulation and associated regional climate change (Xie et al. 2015).

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Table captions

Table 1. Correlation coefficients for decadal precipitation trend for 1979–2012 over the tropics (30°S–30°N) between observations and model simulations

Figure captions

Fig. 1 Decadal trends in summertime sea surface temperature (SST; K 34 yr^{-1}) for 1979–2012. Averages for May to September and November to March are shown in the Northern and Southern Hemispheres, respectively. Three-year running means are applied before calculating linear trends. (a) Observed trend in HadISST. (b) 12-member ensemble means of Atl, (c) IO, and (d) Pac runs, respectively. (e) CMIP5 24-model ensemble mean. Models used are listed in Table S1 in the online supplement

Fig. 2 Similar to Fig. 1b–d but for (a) tAtl and (b) tPac runs, respectively

Fig. 3 Similar to Fig. 1 but for summertime precipitation (mm day^{-1} 34yr^{-1}). (a) Observed trends in GPCP, (b) CMAP, and (c) CRU TS v3.23. Stipples indicate areas with trends that are significant at the 95 % confidence level or higher. Areas surrounded by purple lines indicate land monsoon domains (North American Monsoon; NAM, South American Monsoon; SAM, North African Monsoon; NAM, South African Monsoon; SAM, South Asian Monsoon; SAM, East Asian Monsoon; EAM, Australian Monsoon; AUS). (d) Results of Atl, (e) IO, and (f) Pac runs, respectively. (g) 24-model mean of the CMIP5 multiple model ensemble

Fig. 4 Similar to Fig. 2d–f but for (a) tAtl and (b) tPac runs, respectively

Fig. 5 Decadal trends in summertime precipitation (mm day^{-1} 34yr^{-1}) averaged over the global land monsoon domain (GM), Northern and Southern Hemisphere monsoon domains (NH and SH), and individual monsoon domains (purple areas shown in Fig. 3). (a) Observed trends in CRU TS v3.23 (grey bar with solid line), GPCP (dotted line) and CMAP (dashed line). Three error bars indicate 95% confidence intervals for (left to right) CRU, GPCP and CMAP, respectively. (b) Modeled trends in Atl (blue bar), IO (orange bar) and Pac (red bar) runs. Black rectangles represent results of tAtl and tPac runs. Error bars indicate 95% confidence intervals. Grey and black error bars on the blue (red) bars indicate Atl and tAtl (Pac and tPac) runs, respectively. Solid, dotted and dashed grey rectangles are identical to (a)

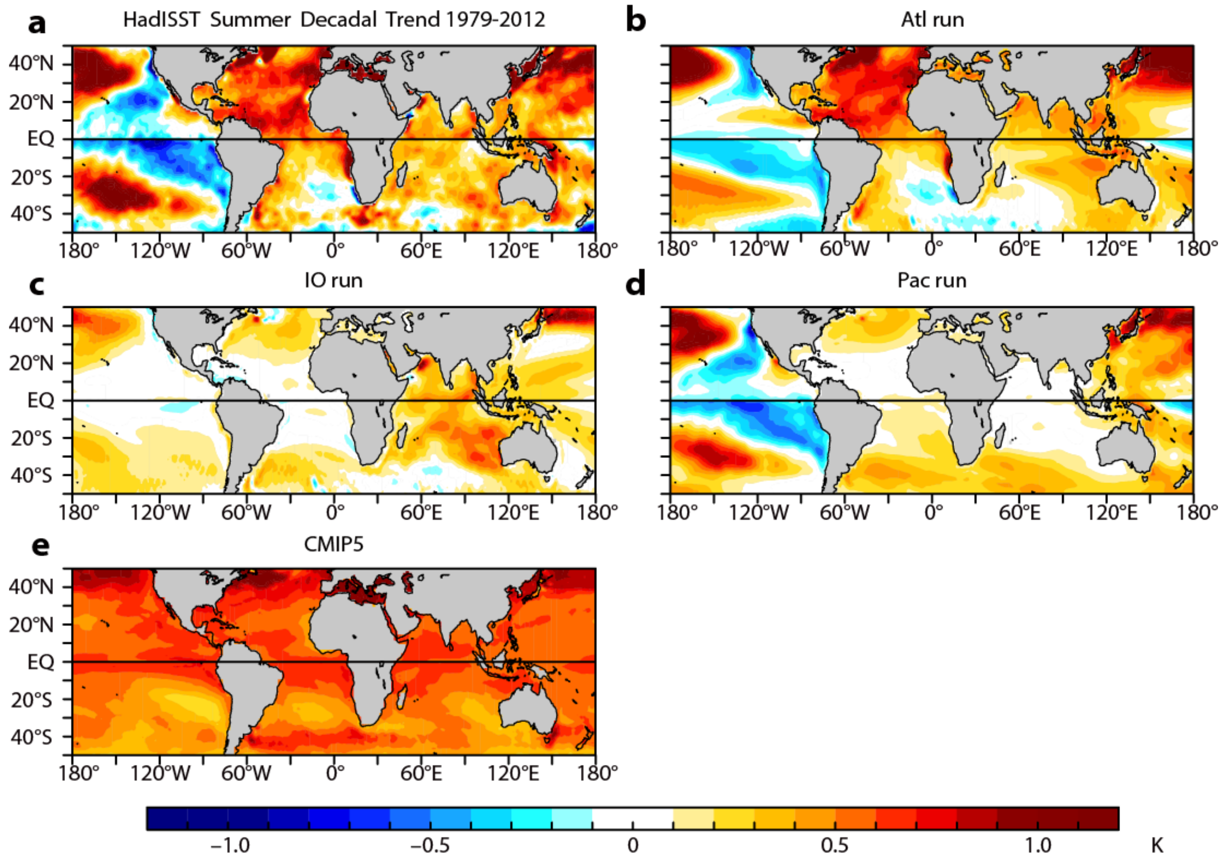
Fig. 6 Summertime mid-to-upper tropospheric thickness (Z200 minus Z500; shading; m) and vertical wind shear (850 hPa minus 200 hPa; vector; m s^{-1}). (a) Climatology for 1979–2012 in ERA-Interim. Thick black lines indicate latitudes with peak thickness in the individual hemispheres. (b) Decadal trend for 1979–2012 in ERA-Interim. Dashed lines represent the peak thickness in climatology, identical to the solid lines in (a). (c) Results of Atl, (d) IO, and (e) Pac runs, respectively

Fig. 7 Similar to Fig. 6c–e but for (a) tAtl and (b) tPac runs, respectively

Fig. 8 Difference in decadal trends of summertime mid-to-upper tropospheric thickness (Z200 minus Z500) for 1979–2012 (m $34yr^{-1}$) between the peak latitude (black line in Fig. 6) and the equator. (a) ERA-Interim (grey), Atl (blue), IO (orange), and Pac (red) runs over the Northern and (b) Southern Hemispheres, respectively. Shadings represent 95% confidence intervals. Dashed blue and red lines indicate tAtl and tPac runs, respectively. Black lines in the lower parts represent longitudinal areas of the land monsoon domains

Table 1. Correlation coefficients for decadal precipitation trend for 1979–2012 over the tropics (30°S–30°N) between observations and model simulations

Experiments	GPCP	CMAP
Atl	0.26	0.30
IO	0.11	–0.02
Pac	0.33	0.15
tAtl	0.26	0.27
tPac	0.32	0.14
CMIP5	0.24	–0.10



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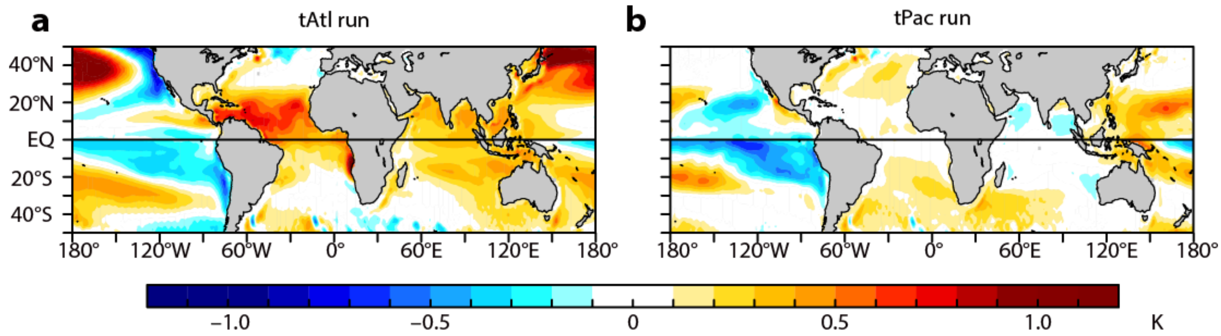
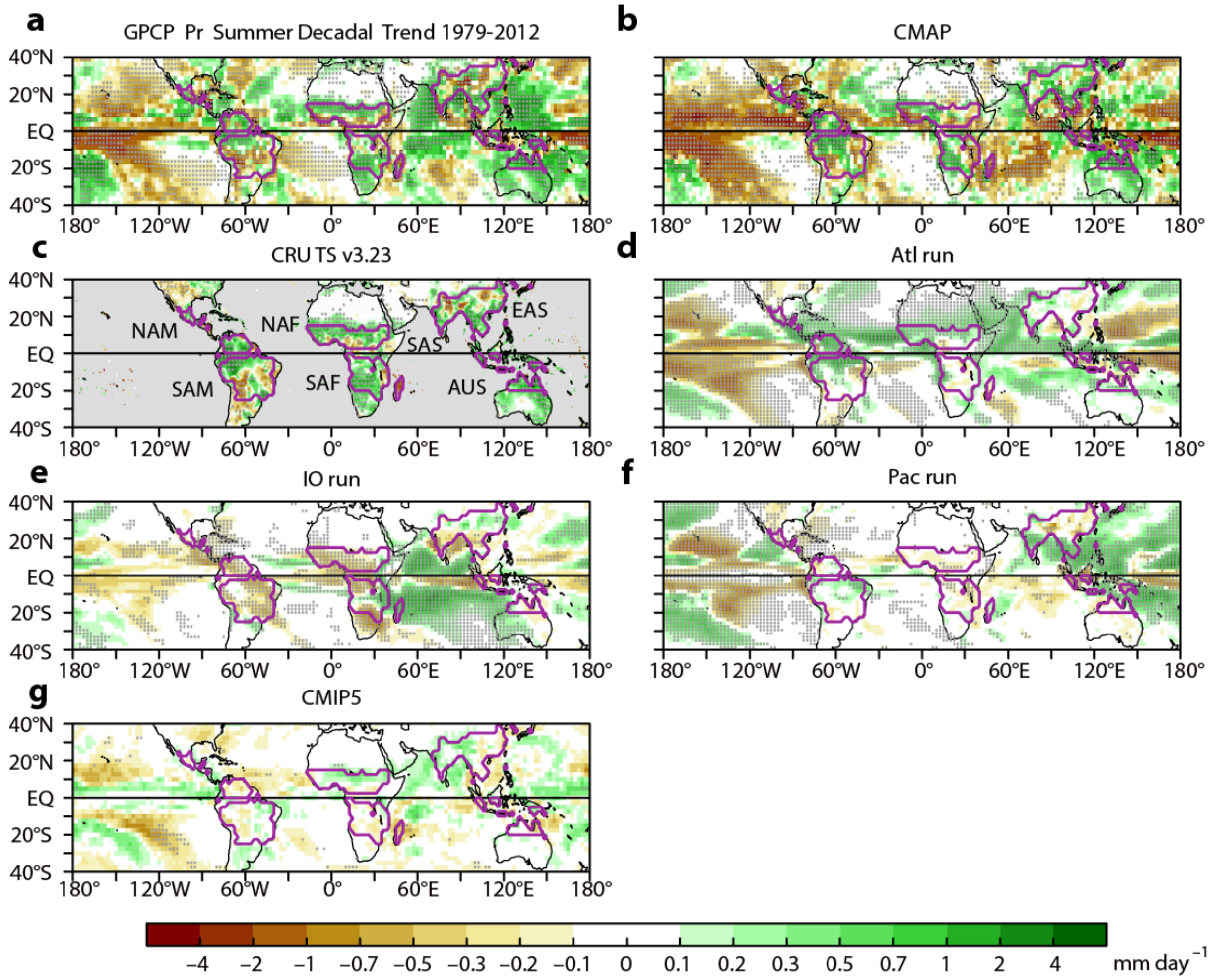


Fig. 2 Similar to Fig. 1b–d but for (a) tAtl and (b) tPac runs, respectively



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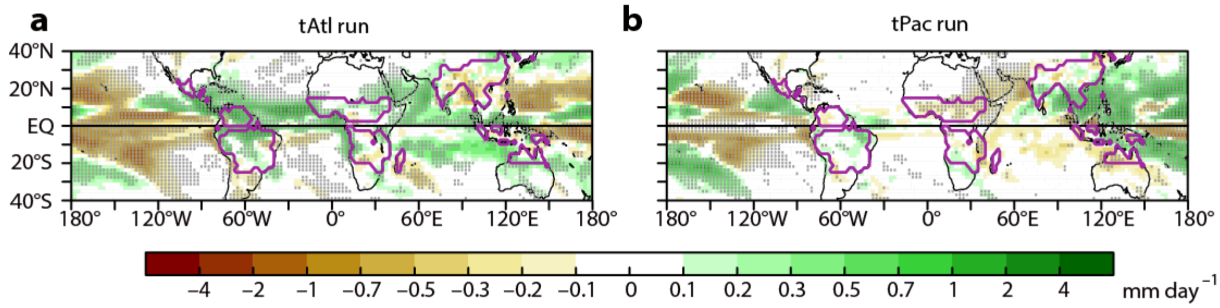

Fig. 4 Similar to Fig. 2d–f but for (a) tAtl and (b) tPac runs, respectively

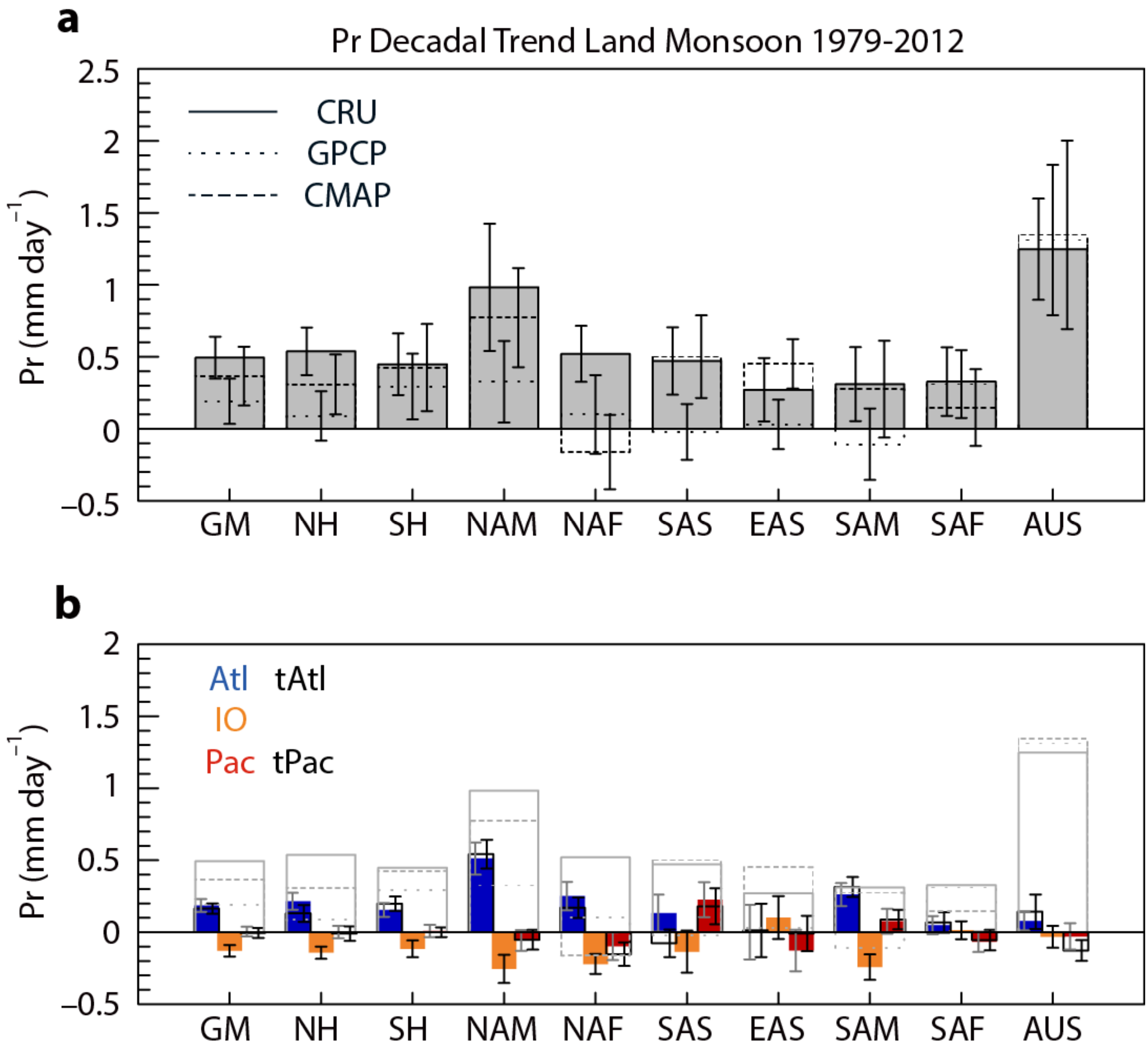

Fig. 5 Decadal trends in summertime precipitation (mm day^{-1} 34yr^{-1}) averaged over the global land monsoon domain (GM), Northern and Southern Hemisphere monsoon domains (NH and SH), and individual monsoon domains (purple areas shown in Fig. 3). (a) Observed trends in CRU TS v3.23 (grey bar with solid line), GPCP (dotted line) and CMAP (dashed line). Three error bars indicate 95% confidence intervals for (left to right) CRU, GPCP and CMAP, respectively. (b) Modeled trends in Atl (blue bar), IO (orange bar) and Pac (red bar) runs. Black rectangles represent results of tAtl and tPac runs. Error bars indicate 95% confidence intervals. Grey and black error bars on the blue (red) bars indicate Atl and tAtl (Pac and tPac) runs, respectively. Solid, dotted and dashed grey rectangles are identical to (a)

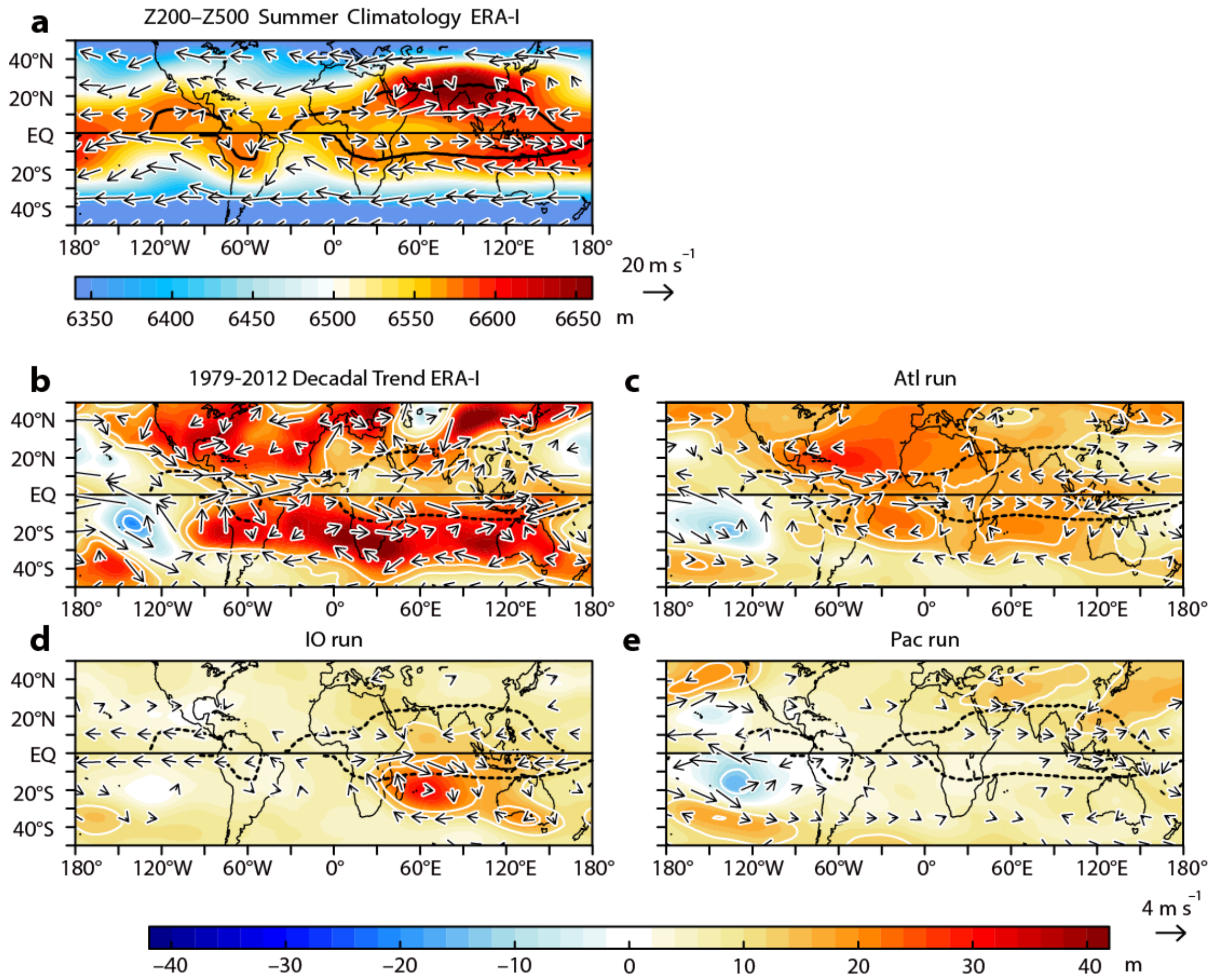

Fig. 6 Summertime mid-to-upper tropospheric thickness (Z200 minus Z500; shading; m) and vertical wind shear (850 hPa minus 200 hPa; vector; m s^{-1}). (a) Climatology for 1979–2012 in ERA-Interim. Thick black lines indicate latitudes with peak thickness in the individual hemispheres. (b) Decadal trend for 1979–2012 in ERA-Interim. Dashed lines represent the peak thickness in climatology, identical to the solid lines in (a). (c) Results of Atl, (d) IO, and (e) Pac runs, respectively

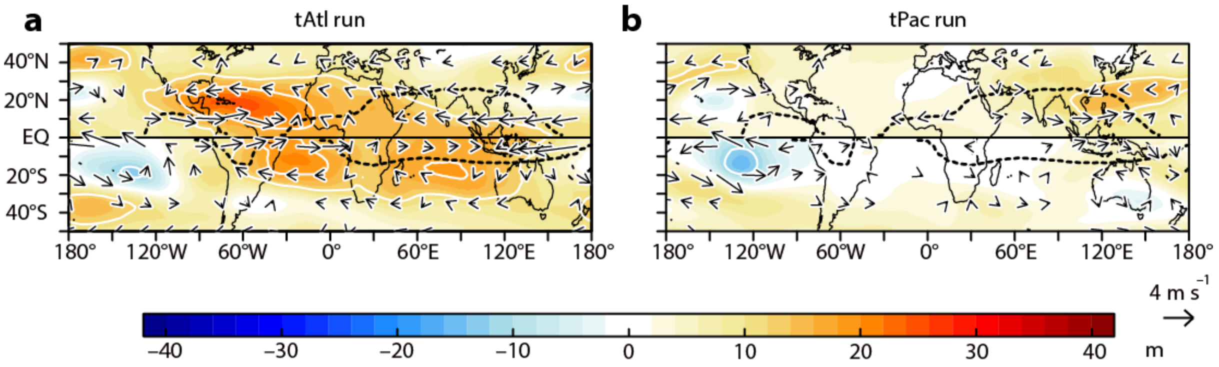

Fig. 7 Similar to Fig. 6c–e but for (a) tAtl and (b) tPac runs, respectively

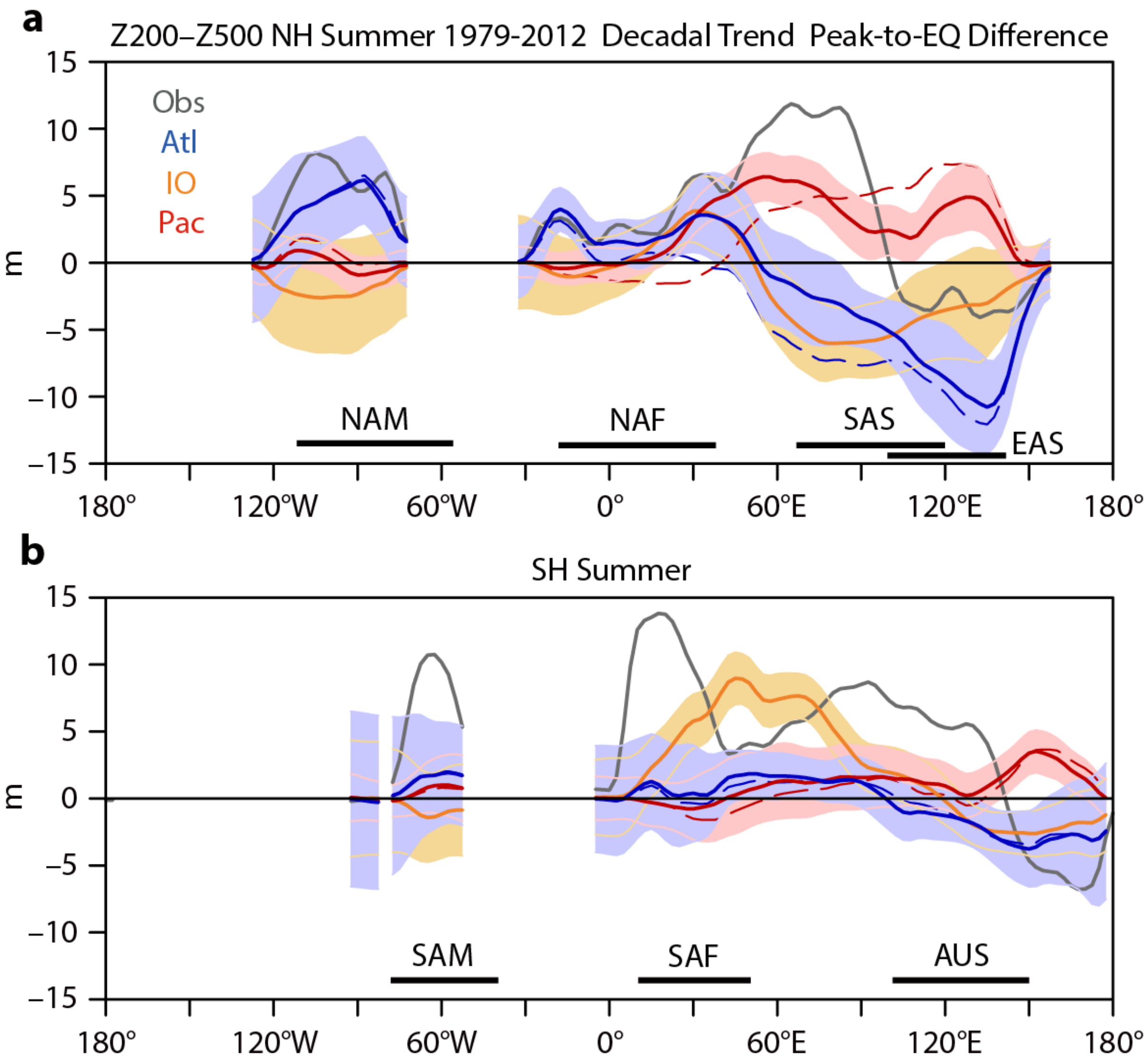

Fig. 8 Difference in decadal trends of summertime mid-to-upper tropospheric thickness (Z200 minus Z500) for 1979–2012 (m $34yr^{-1}$) between the peak latitude (black line in Fig. 6) and the equator. (a) ERA-Interim (grey), Atl (blue), IO (orange), and Pac (red) runs over the Northern and (b) Southern Hemispheres, respectively. Shadings represent 95% confidence intervals. Dashed blue and red lines indicate tAtl and tPac runs, respectively. Black lines in the lower parts represent longitudinal areas of the land monsoon domains