Vol. 72: 1–18, 2017
https://doi.org/10.3354/cr01442

CLIMATE RESEARCH
Clim Res

Published online March 15

Major weather regime changes over Southeast Asia in a near-term future scenario

Cindy Bruyère[2,3], Chainarong Raktham[1,4,*], James Done[2], Jiemjai Kreasuwun[1], Chitrlada Thongbai[1], Wonchai Promnopas[1]

[1]Department of Physics and Materials Science, Faculty of Science, Chiang Mai University, Chiang Mai, 50200, Thailand
[2]National Center for Atmospheric Research, 1850 Table Mesa Drive, Boulder, Colorado 80305, USA
[3]Environmental Sciences and Management, North-West University, Postchefstroom 2531, South Africa
[4]Faculty of Education, Uttaradit Rajabhat University, Uttaradit, 53000, Thailand

ABSTRACT: A near-term future climate scenario over Southeast Asia is generated using a dynamical downscaling approach, and a process-level understanding of the regional climate change is developed by breaking down regional climate into the major rainfall agents of monsoon flow and tropical cyclone activity. The Weather Research and Forecasting model, driven by a Community Climate System Model simulation under the A2 forcing scenario, is used to simulate current and near-term future climate in the Southeast Asia region. Under current climate conditions the model is able to capture the major climate characteristics of the region including the seasonal cycle in tropical cyclone frequency and monsoon precipitation. A near-term future simulation produces an overall increase in the intensity of precipitation events. The strengthening of the Meiyu front combined with an increase in tropical cyclone frequency contributes to this overall increase in precipitation. Future changes in monsoon timing are greater than historical decadal variability, with future onset delayed and future dissipation arriving earlier, reducing monsoon duration. In addition, a higher proportion of zonally-oriented tropical cyclone tracks and higher landfall risk are predicted. This near-term future scenario would result in heightened impacts on already vulnerable communities.

KEY WORDS: Regional climate · Tropical cyclones · Asian Monsoon

1. INTRODUCTION

Heavier rainfall, more intense storms and intensifying droughts are likely to occur in the future as climate change takes effect (IPCC 2012). These extremes, combined with rising sea levels, will increasingly affect already vulnerable coastal areas. The IPCC report on extreme events (IPCC 2012) warns that developing countries will be particularly affected, due in part to their geography, but more so because they have less capacity to be resilient.

Thailand (as an illustrative example of a vulnerable developing country) is located in the middle of the Southeast Asia (SEA) mainland. Its distinguishing geographical features are mountain ranges in the north, lowland central plains and long coastlines. The drainage system in the central plains supports crop cultivation and provides waterways for transportation. These crops contribute 10% of the gross domestic product and provide employment to nearly half of Thailand's population (Office of Agricultural Economics 2015). The long coastal regions are particularly vulnerable to sea level rise and land-falling tropical cyclones (TCs). The nation's capital Bangkok, home to 15% of the population, is under threat of submersion within as little as 15 yr due to rising seas (Philip 2011).

The climate of this region is strongly linked to the onset and dissipation of the Southeast Asian monsoon (Wahl & Morrill 2010, Ummenhofer et al. 2013).

*Corresponding author: glaraktham@gmail.com

Driven by the thermal contrast between oceans and continents and directed by regional topography, the monsoon brings vast amounts of precipitation (Zhou et al. 2011), flash floods and landslides (Matsumoto 1998). In a typical year three-quarters of Southeast Asia rainfall occurs between the middle of May through the middle of October (Lau et al. 1988, Dhar & Nandargi 2003, Meehl & Arblaster 2003, Ueda et al. 2006, Loo et al. 2014). The monsoon trough first moves northward from Thailand through southern China, then moves back southward, signifying the start and end of Thailand's rainy season, and is critically important for agriculture and water resource management. The SEA monsoon exhibits strong interannual variability, resulting in dry years and others marked by excessive flooding (Cook et al. 2010). Understanding and predicting these shifts are essential for flood management and to lessen impacts on lives, property and food production.

TCs also supply significant precipitation to the region (Takahashi et al. 2015). From May through December, Thailand can be affected by TCs originating from both the Bay of Bengal and the Pacific Ocean. The majority of early cyclones originate in the Bay of Bengal, shifting to Pacific storms later in the season (Zhi et al. 2013). Studies (Chen & Yoon 2000, Takahashi & Yasunari 2008, Takahashi et al. 2015) suggest that above average precipitation in the region is directly related to years with enhanced westward-propagating tropical cyclones. Takahashi & Yasunari (2008) estimated that 70% of late season precipitation over Thailand is due to TC activity. The main TC impacts are rainfall-related agricultural impacts in the mountainous northern and northeastern Thailand (Climatological Center, Thai Meteorological Department 2015), another major crop producing area of Thailand (Papademetriou & Dent 2001).

Climate variability and change has the potential to exacerbate monsoon and TC impacts on society. Schewe & Levermann (2012) and Ashfaq et al. (2009) projected significant changes to the onset, duration and impact of the SEA monsoon under climate change scenarios. Thailand is highly vulnerable to flooding and could be impacted significantly if precipitation increases due to climate change (Kreasuwun et al. 2010, Amnuaylojaroen 2014, Maijandee 2014). Information on the potential for near-term change in seasonal rains could therefore support government adaptation and mitigation planning.

Near-term changes in monsoon circulation and regional TC activity, and subsequent impacts through changes in rainfall, are not well understood, and these are the central question addressed by this study. Coarse resolution global climate models (GCMs) fail to capture the regional climate processes needed to accurately simulate SEA regional climate (Sperber et al. 2013, Karmacharya et al. 2015). For example, they perform poorly at simulating the thermal contrast between oceans and continents (Zhou et al. 2011), resulting in GCMs being unable to accurately simulate monsoon onset and associated precipitation. These limitations make it difficult to use GCMs to evaluate climate change effects on SEA precipitation (Hu et al. 2003).

But a prerequisite for studying the effect of climate change on the onset, duration and intensity of the SEA monsoon is to correctly simulate the monsoon under current climate conditions. A number of studies have downscaled GCM data using a variety of regional climate models (RCMs) to assess regional monsoon simulations for the current climate (Dobler & Ahrens 2010, Polanski et al. 2010, Lucas-Picher et al. 2011, Saeed et al. 2012, Dash et al. 2013), leading to improvements over GCMs in terms of regional physical processes, circulation, structure and rainfall patterns. Raju et al. (2015) identified model bias in the monsoon structure simulated by the Weather Research and Forecasting (WRF) model (Skamarock et al. 2008), but overall this model achieved a good simulation of large-scale seasonal changes in surface pressure, winds and rainfall. In keeping with these general findings, Raktham et al. (2015) established a suitable WRF model configuration that captures TC activity and monsoon circulation over SEA.

Studies that have explored future regional change show mixed results (Turner & Annamalai 2012). An ensemble regional study (Kumar et al. 2011) projects increases in monsoon rainfall over India of ~15% by the 2080s relative to the 1970s. Ashfaq et al. (2009) and Lee at al. (2013), on the other hand, project a suppression of the monsoon and a decrease in overall monsoon rainfall. Lee at al. (2013) utilized the Coordinated Regional Climate Downscaling Experiment (CORDEX) simulations (Giorgi et al. 2009) and showed that precipitation projections are strongly linked to the climate scenarios. They reported a 9% increase in precipitation under RCP 8.5 and a marked decrease of 24.8% under RCP 4.5. Raghavan et al. (2012) used future regional climate changes to show increased future streamflow over the Lower Mekong Basin, while Kimoto (2005) reported projections of a weaker monsoon associated with a northeastern shift in the wave trough. Given the complexities of the regional climate processes in the SEA region and the lack of consensus among future change studies, a deeper understanding is needed.

For the above studies looking at far-future projections, scenario uncertainty is likely to dominate the effects of model uncertainty or internal variability (Hawkins & Sutton 2009), whereas for near-term projections (that are the focus of this study) scenario uncertainty is relatively small. Near-term future change is therefore constrained within the bounds of model uncertainty and internal variability.

Using the WRF model, this study develops a process-level understanding of near-term regional change by breaking down SEA regional climate into the major rainfall agents of monsoon and TCs, and explores near-term change in these rainfall agents. The ability of the WRF model to simulate monsoon and TC statistics for long-term simulations of current climate conditions is evaluated. This is followed by an assessment of how these simulations change in a near-term future scenario.

2. EXPERIMENTAL DESIGN AND DATA

2.1. Regional model simulations

Raktham et al. (2015) used WRF to determine the appropriate domain size, location and physics options to capture TC activity and the monsoon circulation over SEA. They concluded that the following model configuration performed well in this region: a domain covering the area 12° S to 42° N and 70° E to approximately 180° E; a horizontal grid spacing of 36 km, with 51 levels in the vertical, up to a height of 10 hPa; and the following physics options: Noah land surface model (Chen & Dudhia 2001), NCAR Community Atmosphere Model (CAM) long- and shortwave radiation (Collins et al. 2004), Yonsei University planetary boundary layer scheme (Hong et al. 2006), Kain-Fritsch cumulus scheme (Kain 2004) and the Lin microphysics scheme (Lin et al. 1983). In this study, we utilize the same model configuration as Raktham et al. (2015) for the current and near-term future simulations.

Three simulations are performed here to investigate near-term future change in monsoon precipitation and TC activity. The first is a 5 yr (2001–2005) current climate simulation driven by reanalysis data, hereafter referred to as 'base climate'; the second is a 10 yr current climate simulation (nominally 1995–2005) using the Community Climate System Model version 3 (CCSM3; Collins et al. 2006) to generate initial and boundary conditions (referred to as 'current climate'); and lastly a near-future 10 yr climate simulation (nominally 2020–2030; referred to as 'future climate') also driven with CCSM3 model data, described in the next subsection.

It is possible that these 10 yr periods could be susceptible to internal variability of the RCM, but the internal variability of regional 10 yr average climate variables are generally small compared to interannual variability (e.g. Lucas-Picher et al. 2008). While Done et al. (2014) showed that internal variability of regional TC activity can be large on annual timescales, Wu et al. (2010) showed internal variability to be small on decadal timescales. Perhaps of greater concern is the extent to which differences between the decadal periods are dominated by climate variability rather than climate change. Using a GCM initial condition ensemble Sriver et al. (2015) showed significant spread in decadal average climate variables among ensemble members on the scale of the USA. Using a similar ensemble approach Deser et al. (2012) found the influence of internal climate variability extends 50 yr or more. The optimal approach of running an ensemble of long-term simulations is computationally prohibitive. Instead, our experiments may be interpreted as a plausible future scenario that includes the effects of climate variability and climate change.

The spin-up period for atmospheric processes is typically in the order of hours (Skamarock 2004). In contrast, the land-surface model (LSM) can take months before it reaches equilibrium. Yang et al. (1995) found a nonlinear dependency between the length of LSM spin-up times and model physics. Karlicky (2013) found that long regional climate simulations are in balance within a year, while Lim at al. (2012) reported that the spin-up period in monsoon areas is as short as 3 mo if the simulation starts before the onset of the monsoon flow. We used a full year spin-up period to ensure well-balanced simulations.

2.2. Global datasets

The driving data used for the base climate simulation are the National Centers for Environmental Prediction / National Center for Atmospheric Research Reanalysis Project (NNRP/NCEP1) data (Kalnay et al. 1996). This dataset provides global-gridded atmospheric information 6 hourly at 00:00, 06:00, 12:00 and 18:00 h UTC on a global 2.5° × 2.5° grid. The sea surface temperature (SST) data used are the Reynolds optimum interpolated (OIv2) analysis (Reynolds 1988, Reynolds & Marsico 1993), available at a temporal resolution of a week and on a horizontal grid of 1° × 1°. This dataset is hosted by the National Oceanic and Atmospheric Administration (NOAA), USA.

The current and future climate simulations are driven with data from a CCSM3 simulation run at T85 (resolutions of ~1.4° for atmosphere and for 1° ocean). CCSM3 is a coupled climate model with components representing the atmosphere, ocean, sea ice, and land surface, as described in detail in Collins et al. (2006). The CCSM3 participated in the Coupled Model Intercomparison Project 3 (CMIP3; Meehl et al. 2007). CCSM3 simulations were initialized in 1950 and run under 20th century emissions, as well as under a number of scenarios from the Special Report on Emission Scenarios (IPCC 2000). For this study we use the 20th century scenario as driving data for the current climate simulation, and the SRES A2 scenario for the future climate simulation. As noted earlier, near future regional climate uncertainty is dominated by model uncertainty and internal variability (Hawkins & Sutton 2009) so the choice of SRES scenario is somewhat arbitrary for this study.

GCMs are known to contain systematic errors (biases). These are removed from the CCSM3 fields following Bruyère et al. (2014), and these bias corrected fields are subsequently used to drive the WRF model. This bias correction method retains the day-to-day weather, climate change and variability from the CCSM3 model, while the base seasonally varying climate is provided by the NNRP reanalysis. This limits the adverse effects of bias in the driving data found by Karmacharya et al. (2015).

2.3. Evaluation datasets

The ability of WRF to capture regional observed precipitation patterns is assessed using 2 products. Observed daily Tropical Rainfall Measuring Mission data (TRMM; Huffman et al. 2007) on a 0.25° × 0.25° grid for the period 1998–2014 are provided by the Goddard Distributed Active Archive Center. TRMM is complemented by the longer-period National Centers for Environmental Prediction Climate Forecast System Reanalysis (CFSR; Saha et al. 2010). CFSR is a global, high-resolution, coupled atmosphere–ocean–land–ice system designed to provide the best estimate of the state of the atmosphere. Precipitation is available daily on a 0.5° grid, and the period 1981–2015 is used for evaluation.

TC observations are taken from the International Best Track Archive for Climate Stewardship (IBTrACS; Knapp et al. 2010). IBTrACS is the official dataset for TC best track data (provided by NOAA), and contains globally consistent 6-hourly TC information including location and intensity.

2.4. Tropical cyclone tracking

There are many approaches in the literature for tracking TCs in numerical model data (Walsh et al. 2007). For this study we adopted the Suzuki-Parker (2012) tracking algorithm that has been shown to work well for WRF data at 36 km grid spacing. This algorithm first detects local pressure minima within a 1° radius. Next, all tracks that do not meet the following 3 criteria are removed: (1) the sum of the horizontal temperature difference between the TC and its environment at 700, 500, and 300 hPa is >2 K; (2) the mean 850 hPa wind speed is greater than the mean 300 hPa wind speed; and (3) the 300 hPa horizontal temperature difference between the TC and its environment is greater than that at 850 hPa. Extratropical cyclones are excluded using the Hart (2003) cyclone phase parameters, which specify that cyclone thermal symmetry (B) should be >10, and both lower tropospheric thermal wind (VTL) and upper tropospheric thermal wind (VTU) should be >0. Finally, only TC tracks that maintained surface wind speeds >17 m s^{-1} for ≥2 d are retained.

3. RESULTS

The aim of the present study is to investigate changes to the major rainfall agents in the SEA region in a near-term future scenario. To achieve this the WRF model is first run for the base climate to evaluate the model using our best available estimates of real-world large-scale conditions (i.e. reanalysis). Climate model-driven runs are then performed for current and future climate scenarios. Analysis focuses first on TC activity over the northwest Pacific Ocean and Bay of Bengal and then moves onto monsoon circulation and monsoon precipitation.

3.1. Tropical cyclones: base climate

On average approximately 36 TCs form annually in the northwest Pacific Ocean and the Bay of Bengal, with about two-thirds of these forming during the monsoon season from May through October (IBTrACS; Knapp et al. 2010). Fig. 1 shows the total number of observed and simulated TCs for the period 2001–2005. Since WRF is set up as a free running climate model, a one-to-one correspondence with observed TCs is not expected, but the interannual variability in cyclone numbers should be captured. Taken over the period 2001–2005, WRF (base climate) over-

predicts the annual number of TCs by 20 %. However, for the main TC period (i.e. the monsoon season from May through October), WRF under-predicts TC numbers by 20 % (Fig. 1). This seasonal dependence of TC frequency is investigated in this subsection.

The dominant mode of TC interannual variability in the northwest Pacific is tied to El Niño Southern Oscillation (ENSO) and its impact varies spatially (Chu 2004). To assess the ability of the model to capture interannual variability we conduct an analysis of the northwest Pacific by quadrant by splitting the basin along 20° N in the zonal direction and 140° E in the meridional direction. These quadrants are indicated by 4 boxes in the top panel of Fig. 5, on top of the observed spatial distribution of TC track density for the period 1995–2005. The choice of quadrants was guided by an assessment of historical TC genesis locations according to El Niño and La Niña conditions (Fig. 11.5 in Chu 2004). Chu (2004) showed that El Niño coincides with enhanced TC genesis frequency in the southeast quadrant, and La Niña coincides with increased TC genesis frequency in the northwest quadrant. Given that our period 2001–2005 contains only one El Niño event, in late 2002/early 2003, and no major La Niña event, we focus on the ability of the model to capture interannual variability of TC genesis frequency in the southwest quadrant. This quadrant also benefits from a higher sample size of genesis events. The model performs well in this region (Fig. 1b) with a Pearson's R of 0.94, although caution is required for this correlation based on only 5 data points. A longer period would be required to establish model skill in capturing an ENSO–TC genesis relationship over the northwest Pacific. Correlations are

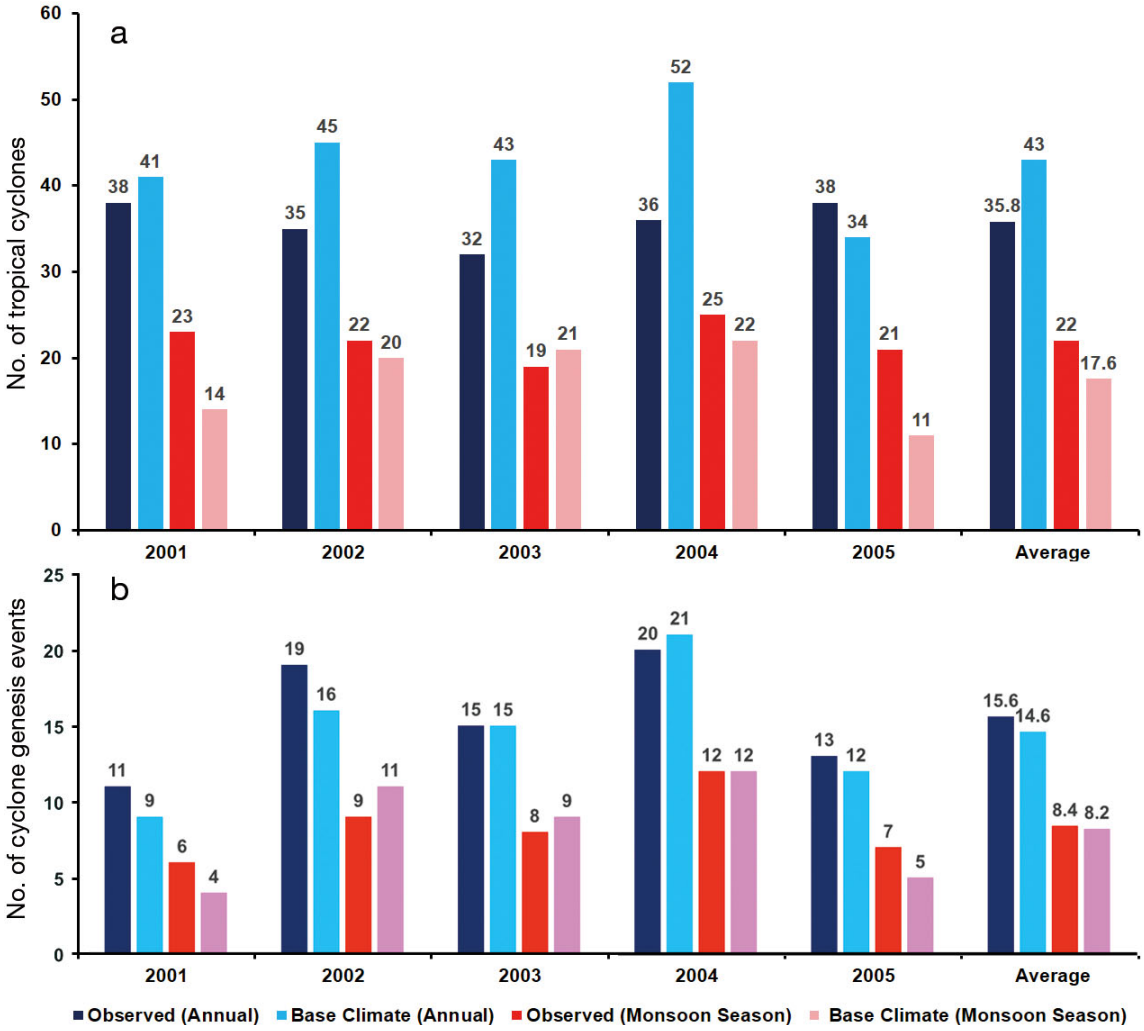

Fig. 1. Frequency of (a) tropical cyclones over the northwest Pacific Ocean and Bay of Bengal, and (b) tropical cyclone genesis events over the southeast quadrant of the northwest Pacific (10° S to 20° N, 140° to 180° E), in the period 2001–2005, showing annual and monsoon season totals based on observed and simulated base climate values

much lower for the other quadrants (not shown) likely due to the weaker ENSO signal.

Fig. 2 shows observed and simulated TC track densities (track density is defined as the number of cyclone tracks within a 5° radius of a point per period) for the period 2001–2005 broken down by annual, monsoon (May to October), and dry (November to April) seasons. The WRF model consistently performs well over the northwest Pacific, in simulating location, track and TC numbers. The model correctly differentiates between the mainly east–west zonally-oriented TC tracks in the dry season, and the east–west and recurving tracks during the monsoon season. The model also performs well in the Bay of Bengal during the monsoon season, but produces an unobserved cluster of cyclones in the Bay of Bengal over South India during the dry season, and it is this erroneous cluster of cyclones that is largely responsible for the over-prediction of annual TC frequency.

Fig. 3 shows the number of observed and simulated (base climate) TCs per month for the Pacific Ocean and the Bay of Bengal, for the period 2001–2005. Observed TCs in the Pacific Ocean occur throughout the year, but generally increase gradually from January, peak in August, then decrease substantially through December, with the bulk of the cyclones forming between May and November. The base climate simulation captures this seasonal cycle well (Fig. 3), although the peak is shifted slightly to later in the season.

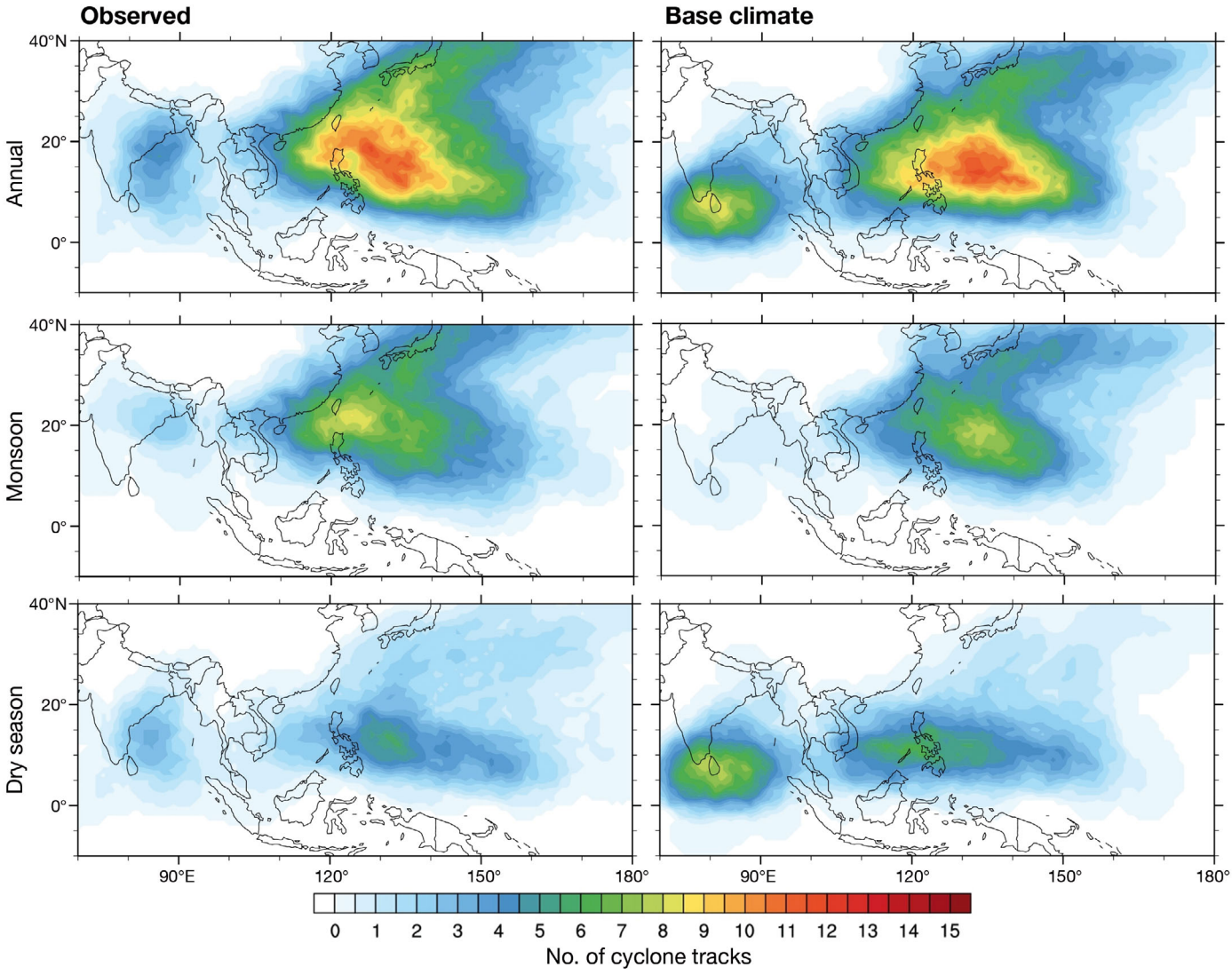

Fig. 2. Tropical cyclone track density in 2001–2005, based on observed (left) and simulated base climate data (right), showing (from top to bottom) annual, monsoon, and dry seasons track densities. Density is defined as the number of cyclone tracks within a 5-degree radius of a point per year

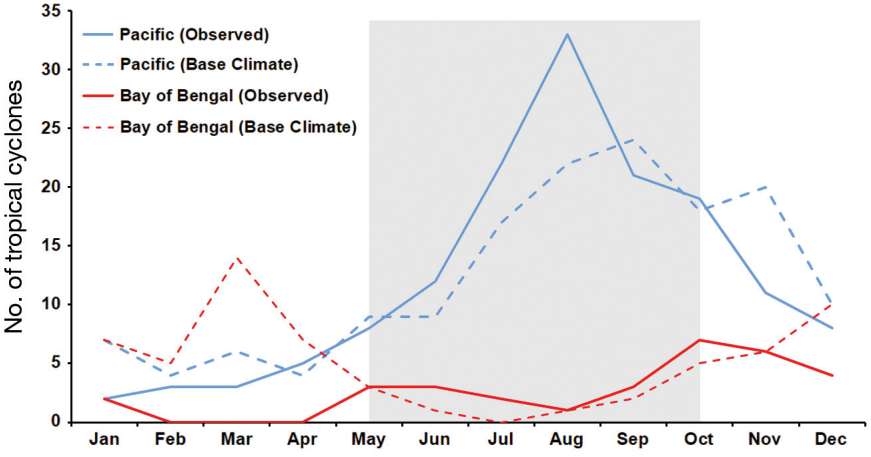

Fig. 3. Monthly average tropical cyclone frequency for the northwest Pacific and Bay of Bengal in the period 2001–2005, based on observed and simulated base climate data. The grey shaded area indicates the monsoon season

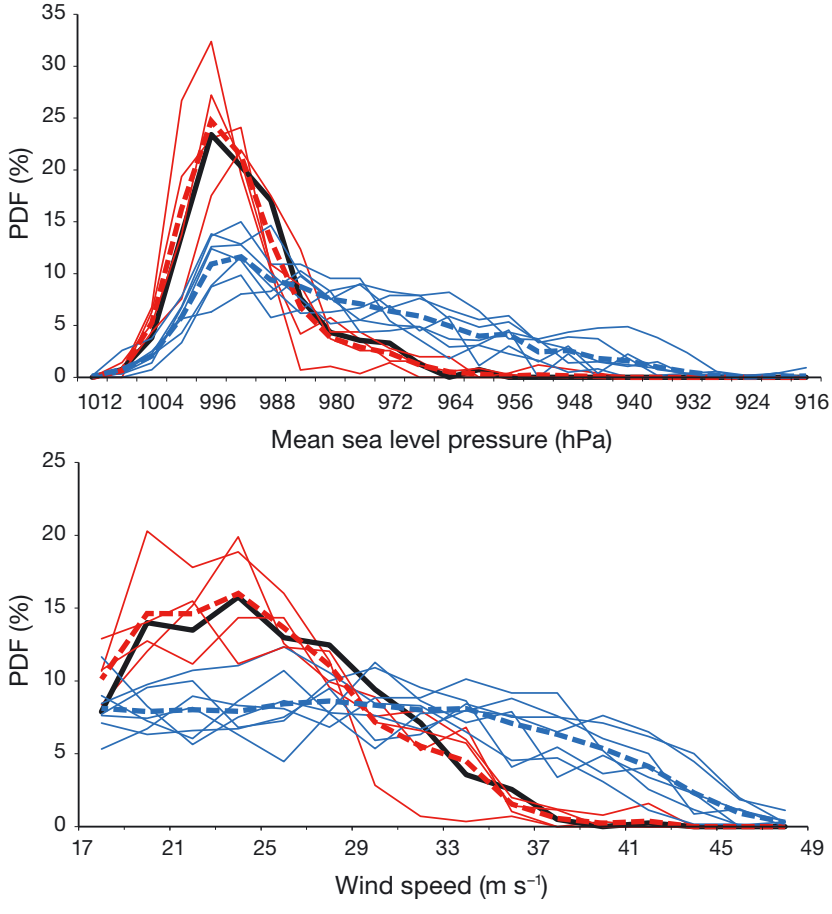

Fig. 4. Probability density functions (PDFs) for mean sea level pressure (top), and maximum wind speed (bottom) from 6-hourly data along the tracks of tropical cyclones in the northwest Pacific and Bay of Bengal, based on simulated base climate data for the monsoon season (May–November, blue) and the dry season (December–April, red). Solid lines: individual months; dashed lines: seasonal averages. Black line: data for the month of March

The Bay of Bengal has significantly fewer TCs than the Pacific Ocean (Fig. 3). During the monsoon season (May to October; grey shading in Fig. 3), there is a clear bimodal signal in TC numbers with a peak both at the beginning and end of the season. At the start of the dry season TC numbers drop off sharply, with few if any developing between February and April. The base climate simulation correctly captures the monsoon season TC cycle, including the peaks at the start and end of the season. During the dry season the simulation fails to simulate the sharp drop-off at the beginning of the dry season. Furthermore, the simulation produces a large number of TCs off South India (Fig. 2) during March, a month that is typically devoid of TC activity. These cyclones, as demonstrated later, are low-pressure cyclonic disturbances that decay as the monsoon circulation establishes.

To understand the potential role of the tracking criteria (Section 2.4) in identifying these erroneous cyclones, we examine the differences between the monsoon and dry season TCs. Fig. 4 shows monthly probability density functions for sea level pressure (SLP) and maximum wind speed created from 6-hourly data along the Base Climate TC tracks. Monsoon season and dry season averages are shown as dashed lines. Monsoon season TCs are significantly stronger than those that develop during the dry season, consistent with observations (Neumann 1993). Cyclones that form during the dry season are generally weak (Fig. 4), achieving the wind speed criteria of 17 m s^{-1} for only short periods. Fig. 4 shows that few, if any, TC wind speeds exceed 36 m s^{-1} during the dry season. Although we do not expect to detect cyclones of Category 5 (on the Saffir-Simpson scale) in a 36 km resolution model (Grasso 2000, Skamarock 2004, Knutson et al. 2010, Murakami & Sugi 2010), simulated TCs are significantly

stronger during the monsoon season, with maximum wind speeds well above the minimum of 17 m s^{-1}. Fig. 4 (black lines) also shows that although cyclone numbers peak in March, these cyclones are not different in comparison to cyclones that develop during other dry months (December through April).

Tracking algorithms used in the literature all use similar tracking parameters, but they often vary in their threshold criteria. In particular variability in the wind speed threshold differs greatly (Walsh et al. 2004, Chauvin et al. 2006, Stowasser et al. 2007, Murakami & Wang 2010), thus leading to the question of whether the erroneous Bay of Bengal cyclones are actually an artifact of our chosen wind speed threshold. Sensitivity tests using slightly different wind speed criteria (not shown) did not significantly alter these results.

Since the specification of the tracking criteria is not the reason for the March Bay of Bengal cyclones, another possibility is that the simulated large-scale environment in the simulation plays a role. Environmental conditions necessary, but not sufficient, for cyclone genesis to occur (Palmén 1948, Riehl 1954, Gray 1979) are a warm ocean, midlevel moisture, a conditionally unstable atmosphere, low-level vorticity, and vertical wind shear through a deep atmospheric layer. Balaguru et al. (2014) found that moisture buildup in this region is the primary source of increased numbers of cyclones in the Bay of Bengal. Simulated mid- and upper-atmosphere flow patterns (not shown) reveal that strong northeasterly winds start to advect moisture towards the southwestern corner of the domain (around the equator and west of 90° E) in late February and early March. This moisture does not exit the domain through the western and southern boundaries but rather accumulates, increasing potential intensity (Emanuel 2007), and supports the development of March cyclones in the Bay of Bengal. These cyclones do not appear every year, only in years with high amounts of available mid- and upper-air moisture. As the monsoon flow establishes, the pool of high moisture is advected away from the domain corner, reducing the available moisture and potential intensity, thus switching off the development of tropical cyclones in the Bay of Bengal.

Accurately simulating the regional climate of Southeast Asia is dependent on the accurate simulation of the monsoon flow. Raktham et al. (2015) demonstrated that monsoon simulations are sensitive to the placement of the western domain boundary. Experiments showed that moving the boundary westward reduces the incorrect development of cyclones in the Bay of Bengal (not shown), but in order to not compromise the simulation of the monsoon (and therefore the entire regional climate simulation), the western boundary was not moved for the current climate and future climate experiments, knowing that this does result in unrealistic development of March cyclones in the Bay of Bengal.

In summary, the base climate simulation performs well, but has biases in TC activity and rainfall. Model biases are commonly found to be non-stationary and strongly dependent on temperature (Christensen & Boberg 2012), but quantifying and treating the changing bias is extremely challenging, requiring more than a simple linear correction (Bellprat et al. 2013). That biases change in time does not preclude us from using the model setup to explore possible future changes, but requires careful interpretation of the results in the context of a constant bias assumption between the current and future periods.

3.2. Tropical cyclones: current and future climate

Observed TC tracks vary from year-to-year (data not shown). Cyclones in the northwest Pacific can be grouped into 3 dominant track pathways (Miller et al. 1988, Harr & Elsberry 1991, Elsner & Liu 2003, Camargo et al. 2007). Depending on the specific atmospheric conditions, cyclones tend to either: (1) track mainly east–west; (2) recurve to the north; or (3) develop and remain close to land. All years contain elements of these different track pathways, but one often dominates. The zonally oriented cyclones typically remain south of 20° N, although this band occasionally shifts northwards. The cyclone track density in the Bay of Bengal (not shown) has lower interannual variability in location, with only a slight north–south shift in the track density maxima. The main interannual difference is in cyclone numbers rather than locations. The main track paths are clearly defined when looking at an annual average track density plot (Fig. 5). The observations show a strong tendency for straight-moving cyclones, with a secondary pathway of high density created by recurving cyclones. The cyclones in the Bay of Bengal mainly track west of 90° E and south of 20° N.

The current climate simulation exhibits a much better magnitude of interannual variability of TC numbers for the Bay of Bengal (a normalized standard deviation of 0.49 compared to 0.53 observed) compared to the base climate simulation. For the Pacific however, the magnitude is higher (0.31) than base climate and observations. The TC track pathways in the cur-

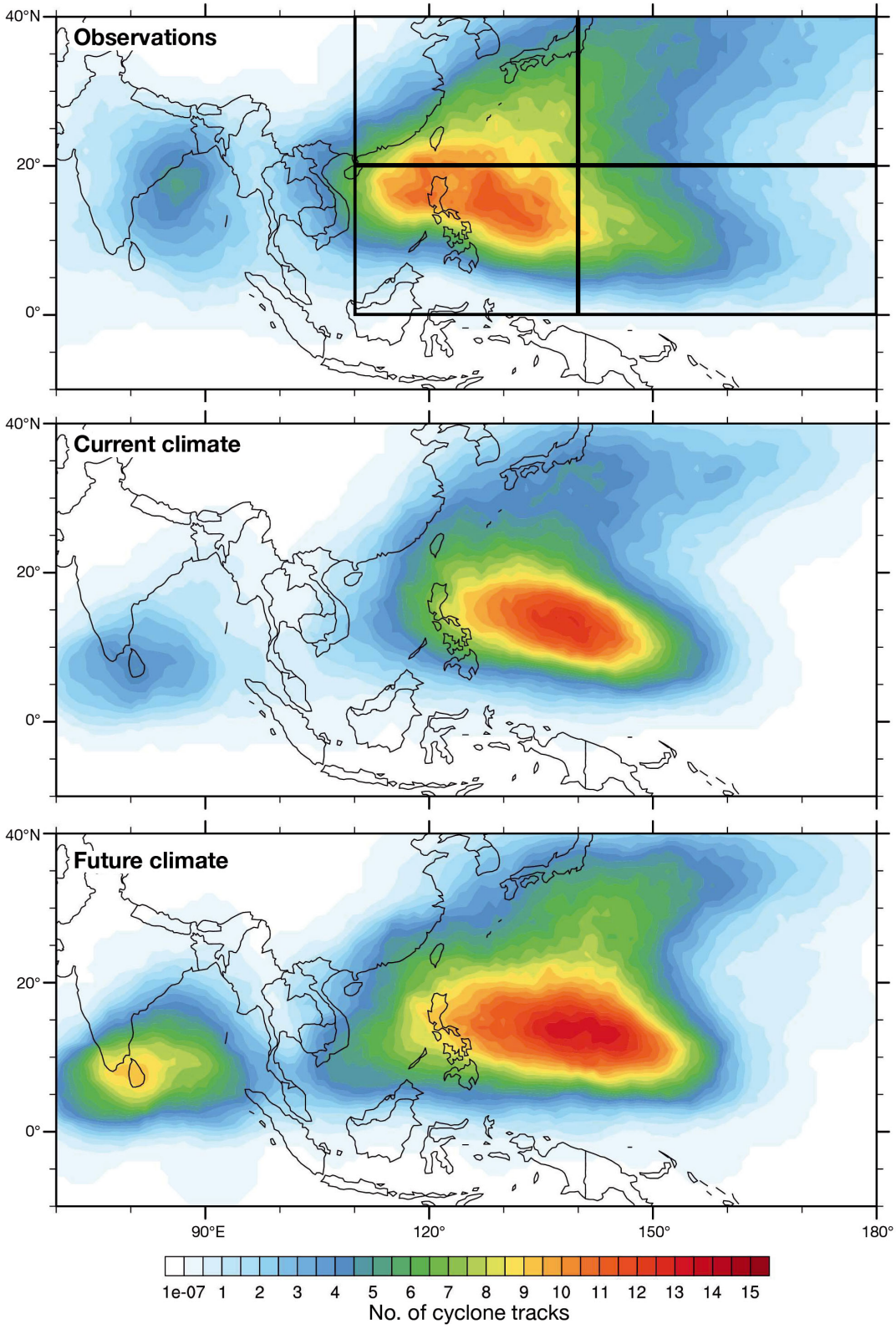

Fig. 5. Annual average tropical cyclone (TC) track density in the northwest Pacific and Bay of Bengal, based on observed data (1995–2005) (top), and simulated data for the current climate (nominally 1995–2005) (middle); and future climate (nominally 2020–2030) (bottom). The boxes in the top panel: quadrants used to evaluate the model's ability to capture the spatial response of TC activity to ENSO

rent climate simulation are also very similar to the observed. Cyclone track density plots (not shown) show that elements of the different observed pathways are present in most years, with one usually dominating. The locations of track density maxima are also generally within the range of observed locations. However, the current climate simulation fails to capture cyclones that develop close to the land (i.e. the third track type over the South China Sea).

The future climate reduces year-to-year variability in TC numbers in both basins. Whereas the observations and the current climate showed similarities, the future climate differs in TC track pathways and frequency. The spatial distribution of tracks is also different, with some years having both zonally-oriented and recurving tracks that dominate; the zonally-oriented pathway become more prominent and more strongly zonal, with more TCs making landfall in the Philippines, Vietnam and Thailand (Fig. 5). Significantly (at the 99% level) more TCs are produced in the future climate compared to the current climate. Fig. 6 shows that the increase in cyclones is mainly in the Pacific Ocean and especially during the monsoon season.

3.3. Precipitation and monsoon

Fig. 7 shows the distinctive annual cycle in observed mean monthly precipitation and sea level pressure patterns. A persistent band of heavy precipitation, associated with an area of low pressure, is present along the equatorial region. During the dry season there is an area of lower pressure in the northeast (around 35° N and 150° E, just off the Japanese coast), which is associated with an area of heavier precipitation. During the monsoon season a rainband associated with the Meiyu front (Lau & Li 1984) is established south of the Pacific subtropical high and, as the subtropical high intensifies, moisture is advected westward resulting in an increase in precipitation over the continental regions. This monsoonal flow, combined with complex topography, results in very high precipitation amounts over the western coastal regions of India, Myanmar and Thailand.

The base climate simulation (Fig. 8) captures this broad annual cycle in precipitation and sea level pressure patterns, but with a wet bias, specifically over the equatorial regions. As discussed in Raktham et al. (2015), this wet bias can be attributed to the Kain-Fritch (Kain 2004) cumulus parameterization scheme. However, we choose to keep this scheme because of its ability to capture reasonable TC activity. Zhu & Shukla (2013) and Zou & Zhou (2013) suggest a role for our imposed one-way air-sea interaction in biases in monsoon flow and associated rainfall. Given the greater potential for model drift using a 2-way interactive ocean (e.g. Ren & Qian 2005), we choose not to further explore the role of the treatment of the air–sea interface and retain focus on potential future changes using one-way coupling.

The current climate simulations (not shown) captured the broad monsoon precipitation and mean sea level pressure patterns, similar to base climate. Figs. 9 & 10 show the future change in precipitation and sea level pressure anomalies (future climate values minus current climate values). under the near-term scenario the pressure over the entire southern half of the domain (Fig. 10) is reduced, associated with higher precipitation amounts in this area (Fig. 9). The Meiyu front intensifies, driving higher precipitation during the peak monsoon months (June, July and August). In addition to the mean precipitation increasing, precipitation increases over all intensity categories, and the most intense precipitation increases to a higher intensity (Fig. 11a). Under this scenario, Thailand in particular experiences substantial increases in mean and extreme precipitation (Figs. 9 & 11b). Thailand already suffers from devastating floods during the monsoon season, and these would be exacerbated under this scenario.

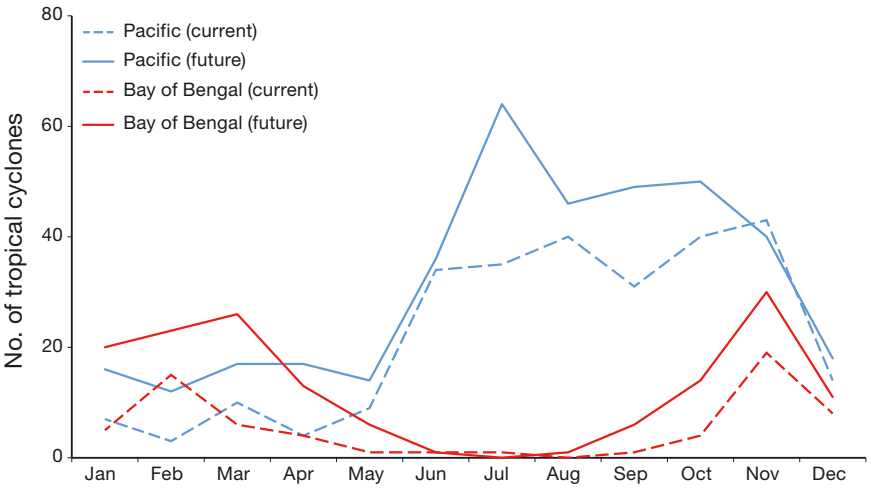

Fig. 6. Monthly average tropical cyclone frequency for the northwest Pacific and Bay of Bengal based on simulated data for the current climate (nominally 1995–2005) and future climate (nominally 2020–2030)

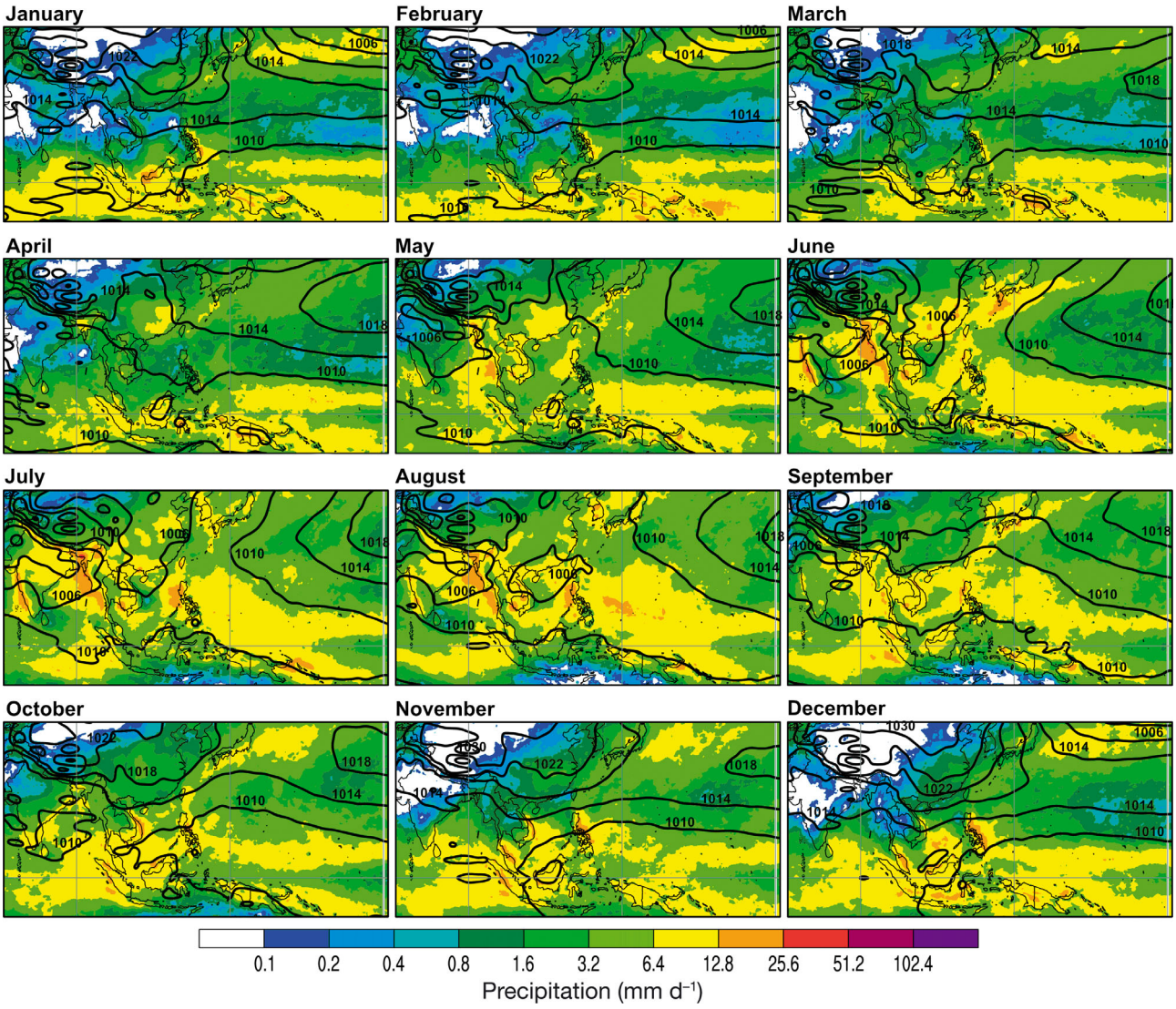

Fig. 7. Observed monthly average precipitation (color) and mean sea level pressure (black contour lines at 4 hPa intervals) in the northwest Pacific and Bay of Bengal, calculated for the period 2001–2005 based on data provided by the Tropical Rainfall Measuring Mission (TRMM) (3B43-gridded data)

3.4. Monsoon onset

Whereas the previous section focused on changes in monsoon precipitation patterns and intensity, the focus here is on monsoon onset, dissipation and duration. Since atmospheric models often have difficulty simulating continental precipitation, using a threshold-based technique to identify monsoon onset could lead to unfair comparison to observations. Here, a proportional threshold is used that normalizes for differences in total rainfall across different datasets. The fractional accumulation method proposed by Sperber et al. (2013) is used as an alternative way to assess monsoon onset in the model simulations.

The fractional accumulation method divides the year into pentads (5 d periods) and pentad average precipitation in the area 12° to 22° N and 95° to 110° E is calculated, accumulated over previous pentads, and normalized by the total. Monsoon onset is defined as the pentad when fractional accumulation exceeds 0.2 and monsoon dissipation is defined as the pentad when fractional accumulation exceeds 0.8. These fractional accumulations create S-curves (Fig. 12) and show a slow and gradual increase in precipitation from the beginning of the year, increasing rapidly between fractions 0.2 and 0.8 (indicating the monsoon season), with a gradual decrease again around October indicating monsoon dissipation.

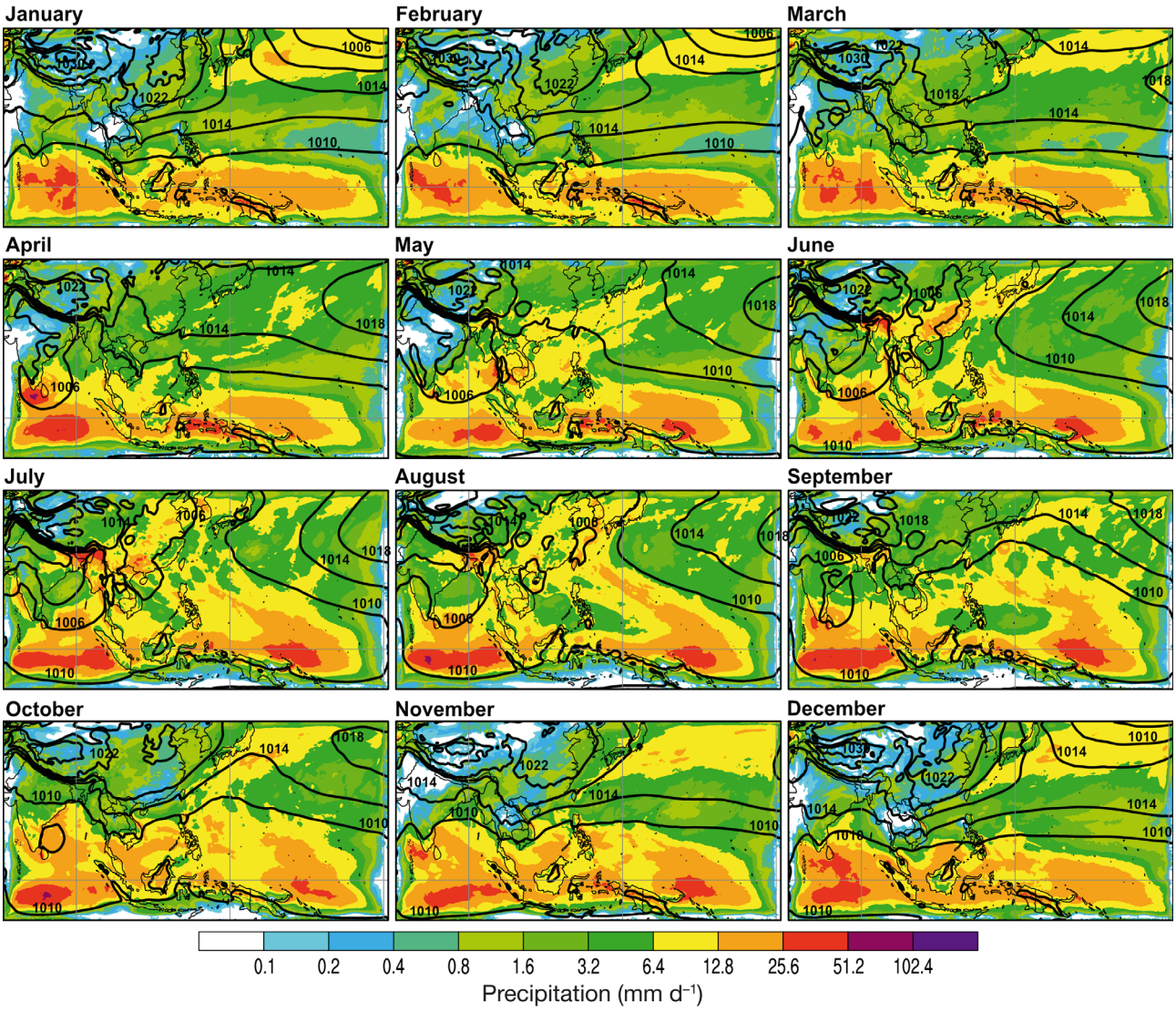

Fig. 8. Monthly average precipitation (color) and mean sea level pressure (black contour lines at 4 hPa intervals) in the northwest Pacific and Bay of Bengal, based on simulated base climate data (2001–2005)

The base climate simulation captures the overall monsoon signal compared to the 2 evaluation datasets (TRMM and CFSR), as shown in Fig. 12. However, monsoon duration is too long, on average, by 4 pentads due to early onset (by 3 pentads, compared to long-term averages from TRMM and CFSR), and late dissipation (on average 1 pentad). Comparisons among base climate, TRMM and CFSR using a consistent period of 2001–2005 show the same base climate model bias (not shown). Monsoon timings in the current climate simulation (not shown) are similar to the base climate.

The observed decadal variability of monsoon onset, dissipation and duration is shown in Fig. 13. Six decades of CFSR data over the period 1981–2015 are defined using overlapping 5 yr periods. Decadal average monsoon dissipation and duration varies by just 1 pentad across the 6 decades, and monsoon onset remains constant at pentad 30. The near-term future scenario shows important changes in monsoon timings (Fig. 13). A future delay in monsoon onset by 2 pentads and a future earlier arrival of monsoon dissipation by 3 pentads means a simulated shorter monsoon in the future climate by 5 pentads. Taking the assumption that model bias is constant in time (as discussed earlier in Section 3.1), the future change may be attributed to either decadal variability or climate change or both. That the near-term future change in monsoon timing lies outside recent historical decadal variability suggests that much of the near-term change may be attributed to climate change.

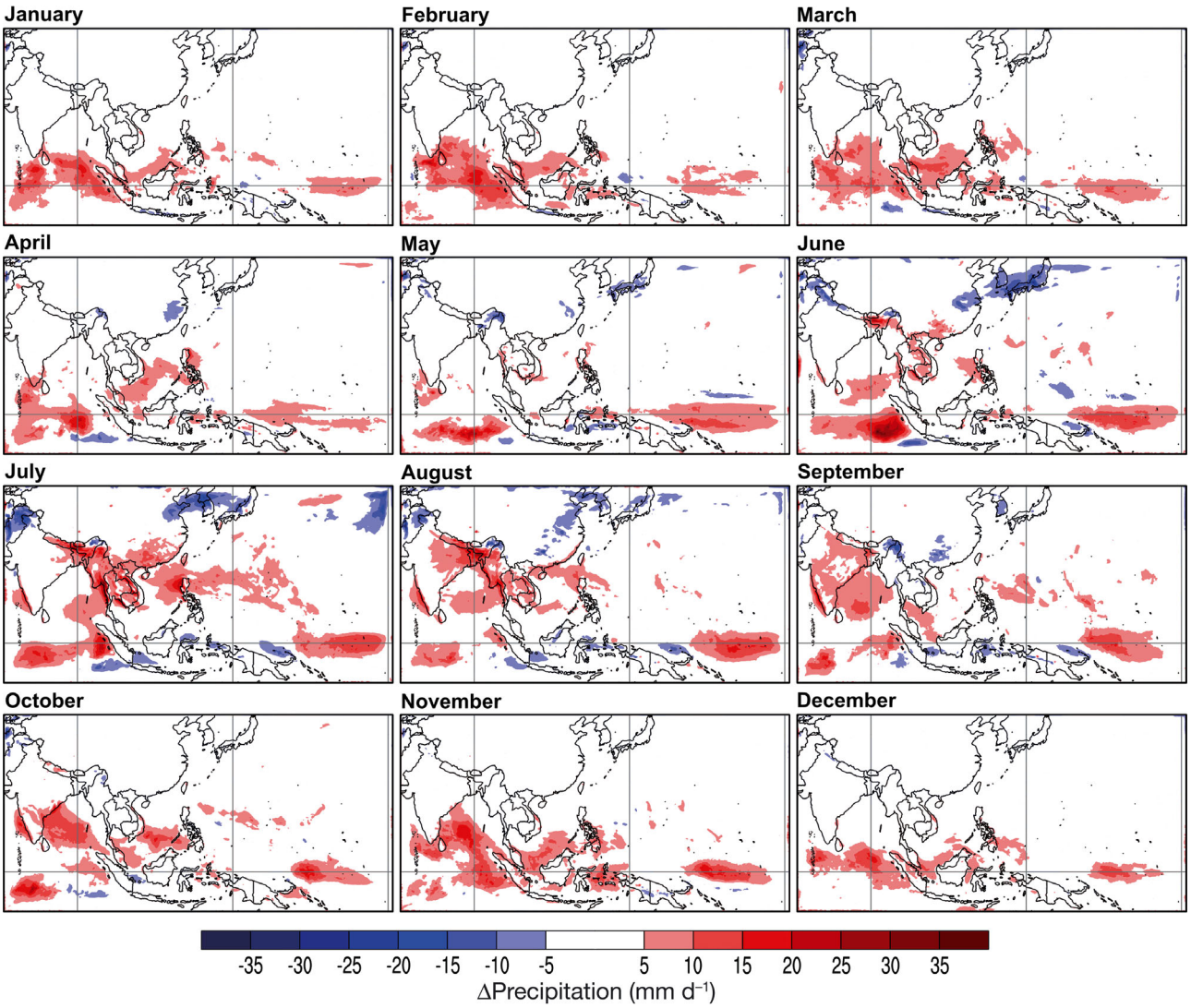

Fig. 9. Future change in monthly average precipitation in the northwest Pacific and Bay of Bengal, based on simulated data, calculated as future climate values minus current climate values

4. DISCUSSION

This section interprets the near-term changes to monsoon rainfall and TC activity found here in the context of previous studies that explored regional climate change.

Uncertainty around this shorter-term outlook is more tightly constrained than for longer-term projections because emissions scenario uncertainty is relatively small on 20 yr timescales. At long timescales, even the sign of the future change to monsoon total rainfall is sensitive to the emissions scenario (Lee et al. 2013). Our near-term scenario shows increases in the mean summer rainfall and increases in extreme rainfall, consistent with long-term changes under a high emissions scenario (Lee et al. 2013). That our near-term scenario shows a shorter monsoon season (due to later onset and earlier dissipation) agrees with other studies that found significant changes at longer timescales. Ashfaq et al. (2009) for example, also found delayed onset in the future. Although our near-term scenario includes both decadal variability and climate change, monsoon timing changes are larger than historical decadal variability. This suggests substantial contributions from climate change for near-term changes.

TC frequency increases in our scenario, mainly in the Pacific and during the monsoon season. In particular, our scenario shows higher risk of landfalling TCs due to increased frequency of zonally orientated tracks across the northwest Pacific. Given that above-average precipitation in the region is directly

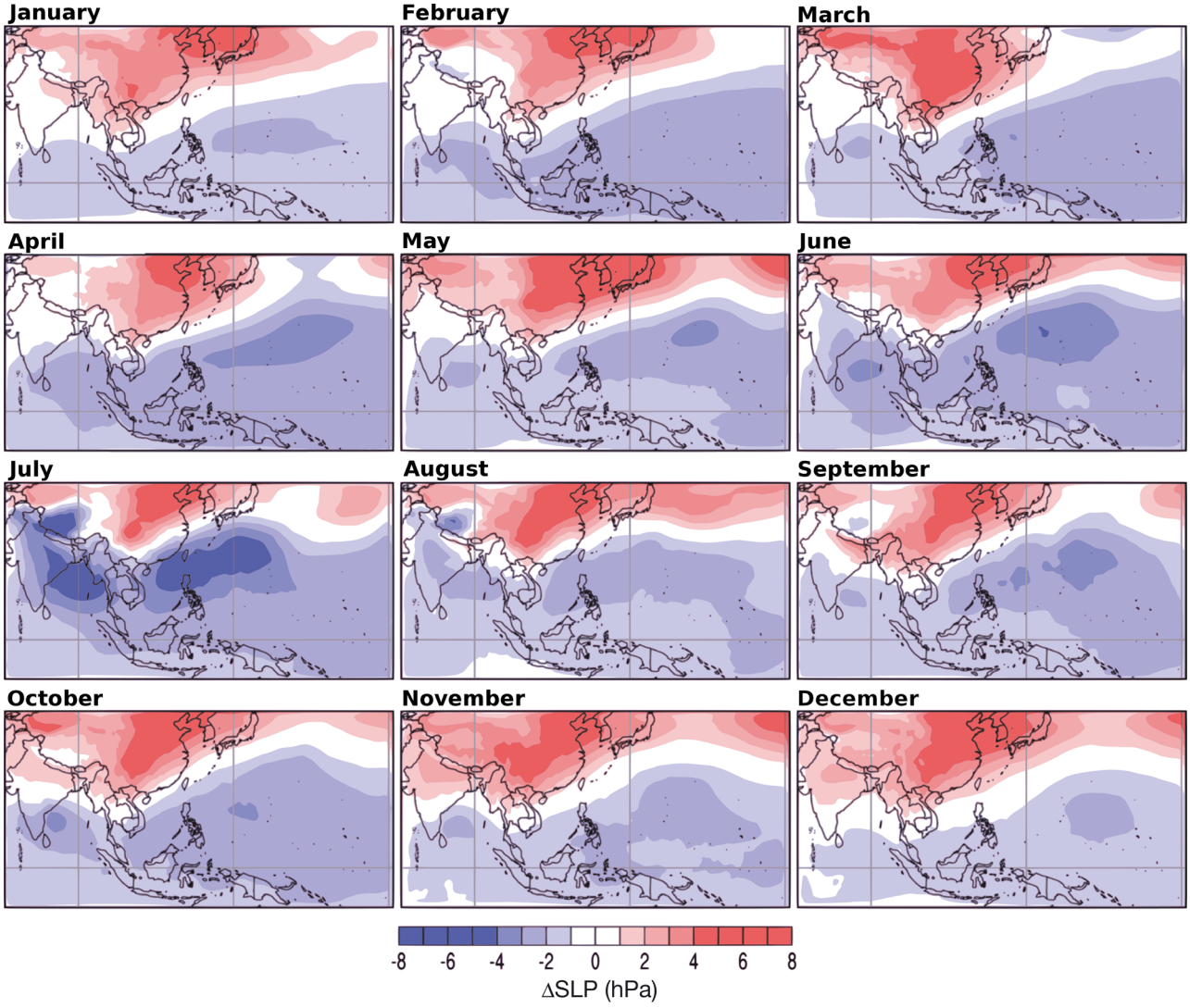

Fig. 10. Future change in monthly mean sea level pressure (SLP) in the northwest Pacific and Bay of Bengal, based on simulated data, calculated as future climate values minus current climate values

related to years with enhanced westward-propagating tropical cyclones (e.g. Takahashi et al. 2015), this future scenario would exacerbate impacts to agriculturally sensitive countries such as Thailand, which already suffers from devastating floods during the monsoon season. Changes to Bay of Bengal TCs are less clear, due to the small sample size and issues with the limited area domain boundary. Previous downscaling studies show the future change in the spatial distribution of TC tracks over the northwest Pacific is highly dependent on the projected large-scale flow (as discussed in Wang & Wu 2015) meaning our simulations, although providing a credible future scenario, are also likely to be dependent on details of the model setup.

This study provides a physically credible near-term scenario under climate variability and change, and did not additionally explore uncertainty about this near-term scenario. The modeling framework outlined here could be applied in an ensemble setting to assess the likelihood of future changes reported here. We also make it clear that this near-term scenario is not a decadal prediction in the sense of using an initialized climate state to predict the phases of the decadal climate variability (Meehl et al. 2009). A decadal prediction study would require additional assessment of predictability of climate variability in the region and its impacts on the major weather regimes.

A number of other limitations include the assumption of constant bias, meaning that the future change is assumed to be unbiased. Further work would be needed to explore the validity in this assumption. Finally, the one-way interactive ocean further con-

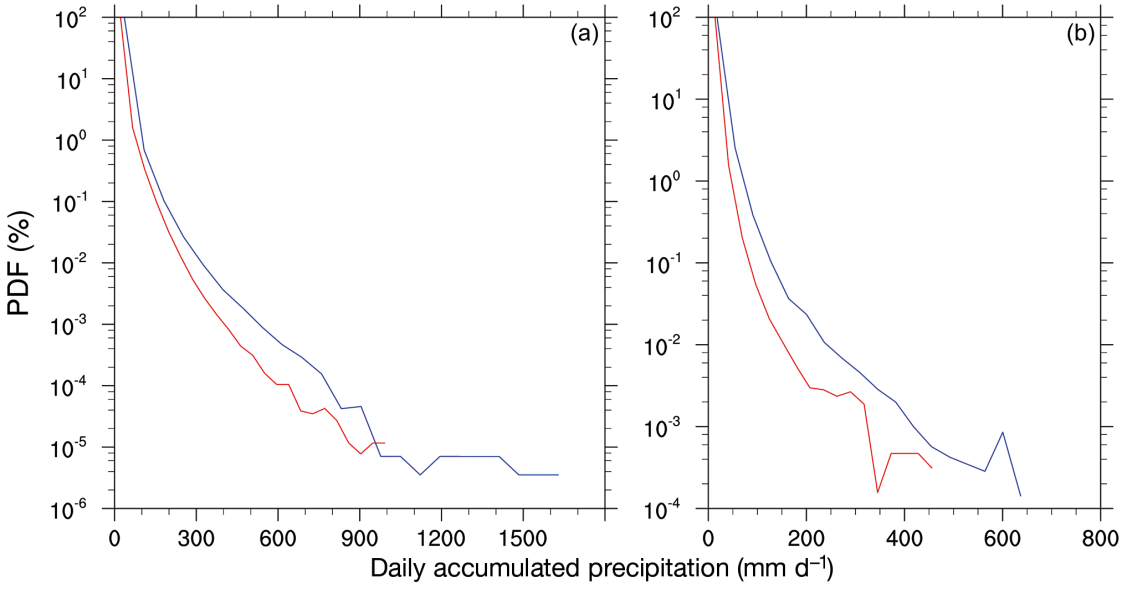

Fig. 11. Probability distribution function (PDF) for monsoon season precipitation (a) over all landmasses in the Weather Research and Forecasting (WRF) simulation domain and (b) over Thailand. Current climate is shown in red and future climate in blue

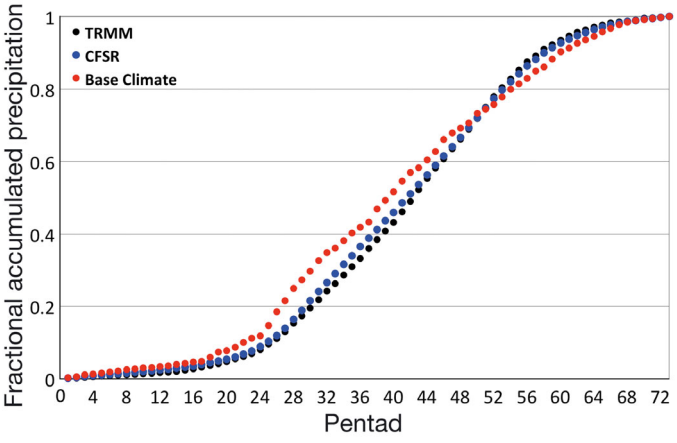

Fig. 12. Fractional accumulated precipitation based on Tropical Rainfall Measuring Mission (TRMM) data (1998–2014), Climate Forecast System Reanalysis (CFSR) data (1981–2015) and the base climate simulation (2001–2005). The onset/dissipation of monsoon is indicated by the 0.2/0.8 fractions

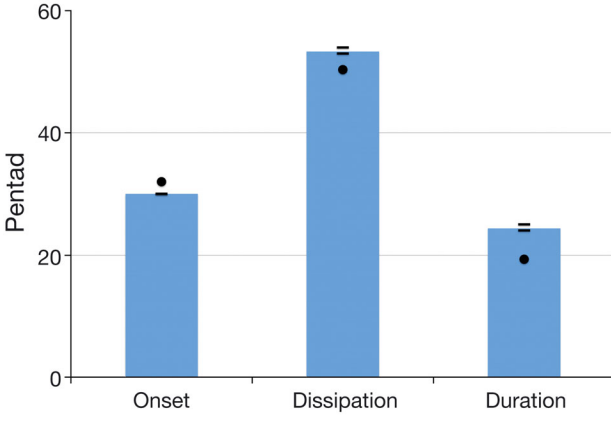

Fig. 13. Monsoon onset, dissipation and duration (pentads). The blue bars indicate the average Climate Forecast System Reanalysis (CFSR) value over the period 1981–2015. Black horizontal bars: minimum and maximum decadal average from CFSR for 6 decades (1981–1990, 1986–1995, 1991–2000, 1996–2005, 2001–2010 and 2006–2015). Black dots: simulated future change (future climate value minus current climate value, where the current climate value has been bias corrected)

strains uncertainty by disallowing the impacts of potentially important coupled processes on future changes.

5. CONCLUSION

This paper assesses potential near-term changes to the major rainfall agents in the Southeast Asia region using a near-term climate scenario with specific focus on TCs and monsoon circulations. A process-based level of understanding was developed using WRF model simulations, using the model configuration of Raktham et al. (2015), driven by a CCSM simulation under A2 forcing scenario to simulate current and near-term future climate (representative of the 2020s) in the Southeast Asia region.

Both the base and current climate simulations are able to capture TC locations, tracks and average numbers. Both simulations correctly differentiate between the mainly east–west zonally-oriented TC

tracks in the dry season, and the recurving tracks during the monsoon season. The simulations also capture the annual pressure and rainfall patterns associated with the monsoon, although with a wet bias. Monsoon timing is generally well captured, but an early monsoon onset and late dissipation result in the simulated monsoon extending longer than observations in both base climate and current climate simulations.

Despite a shorter monsoon season in the near-term future scenario, monsoon total rainfall increases, consistent with the positive rainfall trend found by Kumar et al. (2011). The intensity of rain events and the number of TCs that develop during the monsoon season also increase. In addition, in the near-term scenario a higher proportion of the TC tracks are zonally oriented, therefore also leading to increased landfall risk. Enhanced landfall risk during the monsoon season would lead to heightened impacts on already vulnerable communities.

This study assesses changes in regional weather under one of many possible near-term future climate scenarios. It is likely that some aspects of the changes found here are model dependent, thereby motivating use of a multi-model ensemble in future work to capture a greater range of near-future scenarios.

Acknowledgements. We thank 3 anonymous reviewers for their valuable comments and suggestions to improve the quality of this manuscript. The National Center for Atmospheric Research is sponsored by the National Science Foundation, and this work was partially supported by National Science Foundation grants 1048829 and 1419563, the Willis Research Network, and the Development and Promotion in Science and Technology (DPST) under the Thai government.

LITERATURE CITED

Amnuaylojaroen T (2014) Impacts of projected climate change on tropospheric ozone and PM10 in Thailand. PhD Thesis, Chiang Mai University

Ashfaq M, Shi Y, Tung W, Trapp RJ, Gao X, Pal JS, Diffenbaugh NS (2009) Suppression of South Asian summer monsoon precipitation in the 21st century. Geophys Res Lett 36:L01704

Balaguru K, Taraphdar S, Leung LR, Foltz GR (2014) Increase in the intensity of postmonsoon Bay of Bengal tropical cyclones. Geophys Res Lett 41:3594–3601

Bellprat O, Kotlarski S, Lüthi D, Schär C (2013) Physical constraints for temperature biases in climate models. Geophys Res Lett 40:4042–4047

Bruyère CL, Done JM, Holland GJ, Fredrick S (2014) Bias Corrections of Global Models for Regional Climate Simulations of High-Impact Weather. Clim Dyn 43: 1847–1856

Camargo SJ, Robertson AW, Gaffney SJ, Smyth P, Ghil M (2007) Cluster analysis of typhoon tracks. I. General properties. J Clim 20:3635–3653

Chauvin F, Royer JF, Déqué M (2006) Response of hurricane-type vortices to global warming as simulated by ARPEGE-climate at high resolution. Clim Dyn 27: 377–399

Chen F, Dudhia J (2001) Coupling an advanced land-surface/hydrology model with the Penn State/NCAR MM5 modeling system. I. Model description and implementation. Mon Weather Rev 129:569–585

Chen TC, Yoon JH (2000) Interannual variation in Indochina summer monsoon rainfall: possible mechanism. J Clim 13:1979–1986

Christensen JH, Boberg F (2012) Temperature dependent climate projection deficiencies in CMIP5 models. Geophys Res Lett 39:L24705

Chu PS (2004) ENSO and tropical cyclone activity. In: Murnane RJ, Liu K-B (eds) Hurricanes and typhoons: past, present, and potential. Columbia University Press, New York, NY, p 297–332

Climatological Center, Thai Meteorological Department (2015) The climate of Thailand – Section 11 tropical cyclones in Thailand. Historical Data 1951–2010, https://www.tmd.go.th/en/archive/thailand_climate.pdf

Collins WD, Rasch PJ, Boville BA, Hack JJ and others (2004) Description of the NCAR Community Atmosphere Model (CAM 3.0). NCAR Technical Note NCAR/TN-464+STR, National Centre for Atmospheric Research, Boulder, CO

Collins WD, Bitz CM, Blackmon ML, Bonan GB and others (2006) The Community Climate System Model Version 3 (CCSM). J Clim 19:2122–2143

Cook ER, Anchukaitis KJ, Buckley BM, D'Arrigo RD, Jacoby GC, Wright WE (2010) Asian monsoon failure and megadrought during the last millennium. Science 328: 486–489

Dash SK, Mamgain A, Pattnayak KC, Giorgi F (2013) Spatial and temporal variations in Indian summer monsoon rainfall and temperature: an analysis based on RegCM3 simulations. Pure Appl Geophys 170:655–674

Deser C, Phillips AS, Bourdette V, Teng H (2012) Uncertainty in climate change projections: the role of internal variability. Clim Dyn 38:527–546

Dhar ON, Nandargi S (2003) Hydrometeorological aspects of floods in India. Nat Hazards 28:1–33

Dobler A, Ahrens B (2010) Analysis of the Indian summer monsoon system in the regional climate model COSMO-CLM. J Geophys Res Atmos 115:D16101

Done JM, Bruyère CL, Ge M, Jaye A (2014) Internal variability of North Atlantic tropical cyclones. J Geophys Res Atmos 119:6506–6519

Elsner JB, Liu Kb (2003) Examining the ENSO-typhoon hypothesis. Clim Res 25:43–54

Emanuel KA (2007) Environmental factors affecting tropical cyclone power dissipation. J Clim 20:5497–5509

Giorgi F, Jones C, Asrar GR (2009) Addressing climate information needs at the regional level: the CORDEX framework. WMO Bulletin 58:175–183

Grasso LD (2000) The difference between grid spacing and resolution and their application to numerical modeling. Bull Am Meteorol Soc 81:579–580

Gray WM (1979) Hurricanes: their formation, structure and likely role in the tropical circulation. In: Shaw DB (ed) Meteorology over the tropical oceans. Royal Meteorological Society, Bracknell, p 155–218

Harr PA, Elsberry RL (1991) Tropical cyclone track characteristics as a function of large-scale circulation anom-

alies. Mon Weather Rev 119:1448–1468
Hart RE (2003) A cyclone phase space derived from thermal wind and thermal asymmetry. Mon Weather Rev 131: 585–616
Hawkins E, Sutton R (2009) The potential to narrow uncertainty in regional climate predictions. Bull Am Meteorol Soc 90:1095–1107
Hong SY, Noh Y, Dudhia J (2006) A new vertical diffusion package with an explicit treatment of entrainment processes. Mon Weather Rev 134:2318–2341
Hu ZZ, Yang S, Wu R (2003) Long-term climate variations in China and global warming signals. J Geophys Res 108: 4614
Huffman GJ, Adler RF, Bolvin DT, Gu G and others (2007) The TRMM Multi-satellite Precipitation Analysis: quasi-global, multi-year, combined-sensor precipitation estimates at fine scale. J Hydrometeorol 8:38–55
IPCC (2012) Summary for policymakers. In: Managing the risks of extreme events and disasters to advance climate change adaptation: a special report of Working Groups I and II of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, p 1–19
IPCC (2000) Summary for policymakers: emissions scenarios. A special report of IPCC Working Group III, Intergovernmental Panel on Climate Change, Geneva
Kain JS (2004) The Kain–Fritsch convective parameterization: an update. J Appl Meteorol 43:170–181
Kalnay E, Kanamitsu M, Kistler R, Collins W and others (1996) The NCEP/NCAR 40-year reanalysis project. Bull Amer Meteor Soc 77:437–470,
Karlicky J (2013) Regional climate simulations with WRF Model. In: Šafránková J, Pavl J (eds) Proc 22nd Annu Conf Doctoral Students – WDS 2013, Prague, 4–7 Jun 2013, Part III. MATFYZPRESS, Prague, p 80–84
Karmacharya J, Levine RC, Jones R, Moufouma-Okia W, New M (2015) Sensitivity of systematic biases in South Asian summer monsoon simulations to regional climate model domain size and implications for downscaled regional process studies. Clim Dyn 45:213–231
Kimoto M (2005) Simulated change if the east Asian circulation under global warming scenario. Geophys Res Let 32: L16701
Knapp KR, Kruk MC, Levinson DH, Diamond HJ, Neumann CJ (2010) The International Best Track Archive for Climate Stewardship (IBTrACS): unifying tropical cyclone best track data. Bull Am Meteorol Soc 91:363–376
Knutson TR, McBride JL, Chan J, Emanuel K and others (2010) Tropical cyclone and climate change. Nat Geosci 3:157–163
Kreasuwun J, Chotamornsak C, Ratjiranukul P, Wirunwetchayan O, Amnauylawjarurn T, Phansuke P (2010) Climate change simulation for Thailand by MM5 Regional Climate Model. Final report to Thailand Research Fund (TRF), Bangkok
Kumar KK, Patwardhan SK, Kulkarni A, Kamala K, Rao KK, Jones R (2011) Simulated projections for summer monsoon climate over India by a high-resolution regional climate model (PRECIS). Curr Sci 101:312–326
Lau KM, Li MT (1984) The monsoon of East Asia and its global associations — a survey. Bull Am Meteorol Soc 65: 114–125
Lau KM, Yang GJ, Shen SH (1988) Seasonal and intraseasonal climatology of summer monsoon rainfall over East Asia. Mon Weather Rev 116:18–37
Lee DK, Cha DH, Jin CS, Choi SJ (2013) A regional climate change simulation over East Asia. Asia-Pac J Atmospheric Sci 49:655–664
Lim YJ, Honh J, Lee TY (2012) Spin-up behavior of soil moisture content over East Asia in a land surface model. Meteorol Atmos Phys 118:151–161
Lin YL, Farley RD, Orville HD (1983) Bulk parameterization of the snow field in a cloud model. J Clim Appl Meteorol 22:1065–1092
Loo YY, Billa L, Singh A (2014) Effect of climate change on seasonal monsoon in Asia and its impact on the variability of monsoon rainfall in Southeast Asia. Geoscience Frontiers 6:817–823
Lucas-Picher P, Caya D, de Elia R, Laprise R (2008) Investigation of regional climate models' internal variability with a ten-member ensemble of ten years over a large domain. Clim Dyn 31:927–940
Lucas-Picher P, Christensen JH, Saeed F, Kumar P and others (2011) Can regional climate models represent the Indian monsoon? J Hydrometeorol 12:849–868
Maijandee S (2014) Effects of global warming on regional climate changes over Thailand. PhD Thesis, Chiang Mai University
Matsumoto J (1998) Synoptic features of heavy monsoon rainfall in 1987 related to the severe flood in Bangladesh. Bulletin of the Department of Geography, University of Tokyo 20:43–56
Meehl GA, Arblaster JM (2003) Mechanisms for projected future changes in south Asian monsoon precipitation. Clim Dyn 21:659–675
Meehl GA, Covey C, Taylor KE, Delworth T and others (2007) The WCRP CMIP3 multimodel dataset: a new era in climate change research. Bull Am Meteorol Soc 88: 1383–1394
Meehl GA, Goddard L, Murphy J, Stouffer RJ and others (2009) Decadal prediction: Can it be skillful? Bull Am Meteorol Soc 90:1467–1485
Miller RJ, Tsui TL, Schrader AJ (1988) Climatology of north Pacific tropical cyclone tracks. NAVENVPREDRSCHFAC Tech Rep TR 88-10, Naval Oceanographic and Atmospheric Research Laboratory, Monterey, CA
Murakami H, Sugi M (2010) Effect of model resolution on tropical cyclone climate projects. SOLA 6:73–76
Murakami H, Wang B (2010) Future change of North Atlantic tropical cyclone tracks: projection by a 20-km-mesh global atmospheric model. J Clim 23:2699–2721
Neumann CJ (1993) Global overview. In: Holland GJ (ed) Global guide to tropical cyclone forecasting. WMO/TC-No 560, Report No TCP-31, World Meteorological Organization, Geneva, p 1.1–1.56
Office of Agricultural Economics (2015) Indicator of agricultural economics for Thailand 2015. Ministry of Agriculture and Cooperatives, Thailand, 403:6, http://www.oae.go.th/download/journal/2559/indicator58.pdf (in Thai)
Palmén E (1948) On the distribution of temperature and wind in upper westerlies. J Meteorol 5:20–27
Papademetriou MK, Dent FJ (2001) Crop diversification in the Asia-Pacific region. FAO Regional Office for Asia and the Pacific, Bangkok
Philip B (2011) Thailand, sinking: parts of Bangkok could be underwater in 2030. Time, July 21, 2011 http://content.time.com/time/world/article/0,8599,2084358,00.html
Polanski S, Rinke A, Dethloff K (2010) Validation of the HIRHAM-simulated Indian summer monsoon circulation. Adv Meteorol 2010:415632

Raghavan SV, Vu MT, Liong SY (2012) Assessment of future stream flow over the Sesan catchment of the Lower Mekong Basin in Vietnam. Hydrol Processes 26: 3661–3668

Raju A, Parekh A, Chowdary JS, Gnanaseelan C (2015) Assessment of the Indian summer monsoon in the WRF regional climate model. Clim Dyn 44:3077–3100

Raktham C, Bruyère CL, Kreasuwun J, Done JM, Thongbai C, Promnopas W (2015) Simulation sensitivities of the major weather regimes of the Southeast Asia region. Clim Dyn 44:1403–1417

Ren X, Qian (2005) A coupled regional air–sea model, its performance and climate drift in simulation of the East Asian summer monsoon in 1998. J Clim 25:679–692

Reynolds RW (1988) A real-time global sea surface temperature analysis. J Clim 1:75–86

Reynolds RW, Marsico DC (1993) An improved real-time global sea surface temperature analysis. J Clim 6:114–119

Riehl H (1954) Tropical meteorology. McGraw-Hill, New York, NY

Saeed F, Hagemann S, Jacob D (2012) A framework for the evaluation of the South Asian summer monsoon in a regional climate model applied to REMO. Int J Climatol 32:430–440

Saha S, Moorthi S, Pan HL, Wu X and others (2010) The NCEP climate forecast system reanalysis. Bull Am Meteorol Soc 91:1015–1057

Schewe J, Levermann A (2012) A statistically predictive model for future monsoon failure in India. Environ Res Lett 7:044023

Skamarock WC (2004) Evaluating mesoscale NWP models using kinetic energy spectra. Mon Weather Rev 132: 3019–3032

Skamarock WC, Klemp JB, Dudhia J, Gill DO and others (2008) A description of the advanced research WRF Version 3. NCAR Technical Note TN-475+STR, National Centre for Atmospheric Research, Boulder, CO

Sperber KR, Annamalai H, Kang I, Kitoh A and others (2013) The Asian summer monsoon: an intercomparison of CMIP5 versus CMIP3 simulations of the late 20th century. Clim Dyn 41:2711–2744

Sriver RL, Forest CE, Keller K (2015) Effects of initial conditions uncertainty on regional climate variability: an analysis using a low-resolution CESM ensemble. Geophys Res Lett 42:5468–5476

Stowasser M, Wang Y, Hamilton K (2007) Tropical cyclone changes in the Western North Pacific in a global warming scenario. J Clim 20:2378–2396

Suzuki-Parker A (2012) Uncertainties and limitations in simulating tropical cyclones. Springer Theses, Springer, Berlin

Takahashi HG, Yasunari T (2008) Decreasing trend in rainfall over Indochina during the late summer monsoon: impact of tropical cyclones. J Meteorol Soc Jpn 86:429–438

Takahashi HG, Fujinami H, Yasunari T, Matsumoto J, Baimoung S (2015) Role of tropical cyclones along the monsoon trough in the 2011 Thai flood and interannual variability. J Clim 28:1465–1476

Turner AG, Annamalai H (2012) Climate change and the South Asian summer monsoon. Nat Clim Change 2: 587–595

Ueda H, Iwai A, Kuwako K, Hori ME (2006) Impact of anthropogenic forcing on the Asian summer monsoon as simulated by eight GCMs. Geophys Res Lett 33:L06703,

Ummenhofer CC, D'Arrigo RD, Anchukaitis KJ, Buckley BM, Cook ER (2013) Links between Indo-Pacific climate variability and drought in the Monsoon Asia Drought Atlas. Clim Dyn 40:1319–1334

Wahl ER, Morrill C (2010) Toward understanding and predicting monsoon patterns. Science 328:437–438

Walsh KJE, Nguyen KC, McGregor JL (2004) Fine-resolution regional climate model simulations of the impact of climate change on tropical cyclones near Australia. Clim Dyn 22:47–56

Walsh KJE, Fiorino M, Landsea CW, McInnes KL (2007) Objectively determined resolution-dependent threshold criteria for the detection of tropical cyclones in climate models and reanalysis. J Clim 20:2307–2314

Wang C, Wu L (2015) Influence of future tropical cyclone track changes on their basin-wide intensity over the western North Pacific: downscaled CMIP5 projections. Adv Atmos Sci 32:613–623

Wu L, Tao L, Ding Q (2010) Influence of sea surface warming on environmental factors affecting long-term changes of Atlantic tropical cyclone formation. J Clim 23:5978–5989

Yang Z, Dickinson R, Henderson-Sellers A, Pitman A (1995) Preliminary study of spin-up process in soil–vegetation–atmosphere transfer schemes with the first stage data of PILPS Phase 1a. J Geophys Res 100:16553–16578

Zhi L, Yu W, Li T, Murty VSN, Tangang F (2013) Bimodal character of cyclone climatology in the Bay of Bengal modulated by monsoon seasonal cycle. J Clim 26: 1033–1046

Zhou T, Hsu H-H, Matsumoto J (2011) Summer monsoons in East Asia, Indochina and the western North Pacific. In: Chang CP, Ding Y, Lau NC, Johnson RH, Bin W, Yasunari T (eds) The global monsoon system: research and forecast (2nd edn.). World Scientific Series on Asia-Pacific Weather and Climate. World Scientific Publishing, Singapore 5:43–72

Zhu J, Shukla J (2013) The role of air-sea coupling in seasonal prediction of Asia-Pacific summer monsoon rainfall. J Clim 26:5689–5697

Zou L, Zhou T (2013) Can a regional ocean-atmosphere coupled model improve the simulations of the interannual variability of the western North Pacific summer monsoon? J Clim 26:2353–2367

Editorial responsibility: Filippo Giorgi, Trieste, Italy

Submitted: April 4, 2016; Accepted: November 9, 2016
Proofs received from author(s): February 20, 2017