Future precipitation changes in three key sub-regions of East Asia: the roles of thermodynamics and dynamics

Jiao Li^{1,2} · Yang Zhao^{1,3,4} · Deliang Chen⁵ · Yanzhen Kang⁶ · Hui Wang¹

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Abstract

Previous studies have projected an increase in future summer precipitation across East Asia (EA). This study investigates the relative contributions of thermodynamic and dynamic components to future precipitation changes in three key sub-regions of EA where the maximum centers of the historical precipitation are located (the tropical region, East China, and the Japan and Korea sector), and analyzes the causes of the changes in thermodynamic and dynamic components. Outputs from 30 climate models of the Coupled Model Intercomparison Project Phase 6 (CMIP6) are used. From these, the five best-performing models for historical summer precipitation climatology for EA are selected. The future summer precipitations in the three sub-regions over the near- to mid-term (2020–2069) and the long-term (2070–2095) are then examined using the multi-model ensemble mean of the five models selected (MMM05). The projections were driven by four combined scenarios of the Shared Socioeconomic Pathways (SSPs) and forcing levels of the Representative Concentration Pathways (SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5). The results show that long-term precipitations under SSP5-8.5 are greater than those under the other scenarios across all sub-regions. After the 2070s under SSP5-8.5, a marked precipitation intensification is identified in all three sub-regions, but with different rates of increase. The projected precipitation increase is primarily attributed to the thermodynamic component, while the dynamic component related to circulation changes is relatively weak. Further analysis indicates that the pattern of the thermodynamic component in the three sub-regions is dominated by the climatological upward motion, mediated by an increase in moisture.

Keywords CMIP6 · Sub-regions of East Asia · Future summer precipitation · Thermodynamic component · Dynamic component

1 Introduction

East Asia (EA) is sensitive to variable monsoon precipitation. The East Asian summer monsoon (EASM) is acknowledged to have undergone a decadal shift since the late 1970s, leading to increased drought in the north of the region and flooding in the south (Nitta and Hu 1996; Gong and Ho 2002; Ding et al. 2008, 2010). Droughts and floods have caused great social and economic losses and affected the livelihoods of billions of people in EA (Huang et al. 2007; Zhao et al. 2021). Therefore, reliable projections of future summer precipitation changes in EA are of considerable importance to human well-being and socioeconomic development, including water resource management, agriculture, and ecosystem services.

The global mean surface temperature has increased by 0.85 °C [0.65–1.06 °C] during 1880–2012 (Intergovernmental Panel on Climate Change (IPCC) 2013; Visser

✉ Yang Zhao
zy5000609@snu.ac.kr

✉ Deliang Chen
deliang@gvc.gu.se

¹ State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China

² Institute of Atmospheric Sciences, Fudan University, Shanghai, China

³ Research Institute of Basic Sciences, Seoul National University, Seoul, Republic of Korea

⁴ School of Earth and Environmental Sciences, Seoul National University, Seoul, Republic of Korea

⁵ Regional Climate Group, Department of Earth Sciences, University of Gothenburg, Gothenburg, Sweden

⁶ College of Meteorology and Oceanography, National University of Defense Technology, Changsha, China

et al. 2018). As one of the hotspots of global warming, EA has experienced robust warming in the twentieth century (Hansen et al. 1999, 2006; Guan et al. 2015), particularly after the 1970s (Jeong et al. 2009; Zhu et al. 2012). Moreover, the pronounced global warming is to continue during the twenty-first century (Dickinson and Cicerone 1986; Katarzyna et al. 2020). Li et al. (2018) projected warming of ~ 0.2 °C higher than the global mean in EA during this century. Changes in summer precipitation in EA have been projected as part of the atmospheric response to the warming climate (IPCC 2001; Akio and Takao 2006). Min et al. (2006) projected an increase in EA precipitation in the twenty-first century, at a rate greater than that of the global mean; more recently, Kamae et al. (2014) reported that summer precipitation in EA will strengthen under the projected circulation patterns.

Many studies have advanced our understanding of the physical mechanisms that cause changes in the hydrological cycle in EA under global warming (Lu et al. 2007; Seo and Ok 2013; Li et al. 2019a). Using high-resolution atmospheric models, Woo et al. (2018) reported that the projected summer precipitation increase in EA is caused by enhanced atmospheric moisture along the eastern coast of China and Korea. Using moisture budget analysis, Li et al. (2014) proposed that the projected wetting trend in the Asian monsoon region can be mainly explained by thermodynamic processes (specifically, total mean moisture convergence), but in contrast, the dynamic effect suggests a weakened moisture convergence and a descending motion anomaly in the Asian monsoon region. Further analysis revealed that the thermodynamic component is mainly caused by greenhouse gas forcing, whereas the dynamic component is dominated by aerosol forcing. Endo and Kitoh (2014) identified a positive contribution from the thermodynamic effect in the monsoon region, but a negative contribution from the dynamic effect related to monsoon circulation in the Asian monsoon region, which is weaker than in other monsoon regions. In Northeast Asia, the future increase in summer precipitation under global warming is expected to come from the thermodynamic contribution, but the dynamic component can offset the thermodynamic contribution, thus leading to a decrease in local precipitation in some regions (Lee et al. 2017).

In contrast, Wang et al. (2014) reported that the overall EASM precipitation will increase during the twenty-first century, with the slight enhancement of the subtropical summer monsoon circulation as a result of both the increased zonal thermal contrast and sea level pressure difference. This differs from the above studies, which showed a weakened EASM circulation (Endo and Kitoh 2014; Li et al. 2014). Similarly, Li et al. (2019a) found that both the overall EASM circulation and precipitation will be enhanced under global warming, based on analysis of Phase 6 of the ensemble experiments of the Coupled Model Intercomparison

Project (CMIP6), which is different from the projection for other monsoon regions. Moreover, the relative contributions from the dynamic and thermodynamic components to future precipitation changes are similar when the global warming increase is less than 2 °C, but the dynamic component becomes dominant under further warming. Ha et al. (2020) examined projections in the CMIP6 archive, and diagnosed an earlier onset and later termination of the projected EASM, but their study did not identify the underlying causes of the changes.

All of these previous studies show that EASM precipitation will increase under future warming scenarios. However, the projected EASM circulation changes differ among the various studies. Furthermore, these studies have discussed the possible mechanisms of future summer precipitation changes in response to warming, but are mostly carried out at large spatial scales; for example, the global domain, or the whole of the EA region (Seager et al. 2010; Hsu et al. 2012, 2013; Chen et al. 2020; Moon and Ha 2020; Wang et al. 2020). The changes in regional precipitation, particularly for the EASM, are more complex than changes in global precipitation due to location and other factors relevant to the EASM (Zhai et al. 2005; Wu et al. 2009; Zhao et al. 2018; Li et al. 2019b). Hence, future sub-regional summer precipitation changes in response to warming are far from certain. Therefore, this study aims to explore the following issues: (1) How will the regional precipitation in EA change? (2) Will the dynamic and thermodynamic components play a role in future summer precipitation changes in the individual sub-regions of EA? (3) What are the possible causes underlying the changed thermodynamic and dynamic components? It is imperative to investigate the detailed behavior of future summer precipitation changes and related physical processes across the sub-regions of EA under warming, as part of the process of formulation of countermeasures involving hydrological adaptation and mitigation stemming from specific sub-regional changes to the EASM climate.

The recently released CMIP6 results, organized by the World Climate Research Programme's Working Group, used state-of-the-art climate models from many different groups and institutes, including more comprehensive general coupled models (GCMs) with finer spatial resolution than those in previous phases (Eyring et al. 2016). Several recent studies have analyzed monsoon precipitation in the model simulations from CMIP6 (Zhou et al. 2019, 2020; Chen et al. 2020). To increase confidence in future projections, we first employ a suite of model evaluation statistics to choose the CMIP6 models that achieve reasonably realistic simulations compared to the observed EASM climatology. With these chosen models, we then provide a detailed picture of future summertime precipitation changes in sub-regions of EA and evaluate the relative contributions from dynamic and thermodynamic components.

In the remainder of this paper, Sect. 2 describes the models, data, and methodology. In Sect. 3, diagnostics and skill metrics for the models are calculated to assess the fidelity of CMIP6 models in reproducing the historical summer precipitation characteristics in EA. Section 4 presents the spatial and temporal evolution of future summer precipitation in sub-regions of EA to the end of the twenty-first century. Section 5 investigates the quantitative contributions of dynamic and thermodynamic components to future summer precipitation changes in sub-regions of EA and discusses the possible causes behind the changes in these components. The main conclusions are summarized in Sect. 6.

2 Data and methods

2.1 Models and data

2.1.1 CMIP6 models

There are 30 climate models in the CMIP6 for this study (Eyring et al. 2016). Details of these models are listed in Table 1, including their model number, model acronym, affiliation, and horizontal resolution; further information can be found at <https://esgf-node.llnl.gov/search/cmip6>. Monthly mean data for precipitation, evaporation, near-surface air temperature, specific humidity, vertical velocity, and horizontal wind fields are utilized in this analysis. We focus on the CMIP6 historical simulations and future projections under the four scenarios. The historical simulations are used for model validation; these are performed under anthropogenic (e.g., greenhouse gases and anthropogenic aerosols, land cover and land use change, and ozone) and natural (solar and volcanic activities) forcings (Scoccimarro and Gualdi 2020). The historical simulations for 1981–2014 are used to represent the historical climate. The future scenarios are the combination of the Shared Socioeconomic Pathways (SSPs, O'Neill et al. 2015) and forcing levels of the Representative Concentration Pathways (RCPs). The SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5 experiments are selected, in which the target radiative forcing in the year 2100 reaches 2.6, 4.5, 7.0, and 8.5 W m^{-2} , respectively, relative to the pre-industrial conditions before 1850. Here, the near- to mid-term and long-term refer to 2020–2069 and 2070–2095, respectively (Ahn et al. 2016; Woo et al. 2018). The future changes are acquired by taking the difference between future and historical periods (future minus historical simulations). Although some models have a large number of ensemble members, we use only the first member of each model in this study when computing the multi-model mean (MMM). Previous studies have widely emphasized the substantial biases in precipitation simulations over complex terrain such as the Tibetan Plateau (TP) (Freychet et al. 2015;

Qu 2017); thus, EA here is defined as the area to the east of the main TP (10° – 55° N, 100° – 160° E). The boreal summer season in EA refers to June, July, and August (hereafter referred to as JJA).

2.1.2 Datasets

The most evident features of the EASM are the meridional dipole rainfall pattern and western North Pacific anticyclone (Wang et al. 2008). Thus, precipitation and 850-hPa winds are selected to characterize the EASM. To evaluate the performances of the 30 CMIP6 models, a range of reanalysis datasets, including precipitation and atmospheric circulation variables, need to be compared to ensure consistency with observations. Selecting variables from different reanalysis datasets can help to ensure the robustness of the analysis results (Li and Li 2014). Therefore, multiple datasets are employed, including two observational precipitation datasets: the Global Precipitation Climatology Project (GPCP) monthly precipitation dataset from 1979 to the present that combines observations and satellite precipitation data onto a global $2.5^{\circ} \times 2.5^{\circ}$ grid (Adler et al. 2003) and the Climate Prediction Center Merged Analysis of Precipitation (CMAP) monthly global gridded precipitation dataset from 1979 to the present, with a spatial resolution of $2.5^{\circ} \times 2.5^{\circ}$, which is compiled from rain-gauge observations and satellite estimates (Xie and Arkin 1997). Two atmospheric circulation reanalysis datasets are also used: the National Centers for Environment Prediction (NCEP)-U.S. Department of Energy Intercomparison Project II reanalysis (NCEP-2) from 1979 to the present with a $2.5^{\circ} \times 2.5^{\circ}$ horizontal resolution (Kanamitsu et al. 2002); and the newly released European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA5) from 1979 to near real-time with a spatial resolution of $1.5^{\circ} \times 1.5^{\circ}$ (Hersbach et al. 2020). All data during the period 1981–2014 are extracted from the above datasets. These reanalysis datasets, together with the outputs of the CMIP6 models, are interpolated onto a uniform grid ($1.5^{\circ} \times 1.5^{\circ}$) using bilinear interpolation before the analysis described below.

2.2 Method

2.2.1 Skill score method

For the validation of CMIP6 climate models, the spatial pattern of summer precipitation climatology is considered. Certain metrics have been employed to quantitatively evaluate the 30 climate models and to determine which models best capture the spatial pattern of the EASM precipitation climatology based on the comparison with observations (Taylor 2001; Hirota et al. 2011; Chen et al. 2013). Here we use:

Table 1 Acronyms, institutions, and horizontal resolution of CMIP6 climate models used in this study

No	Model acronym	Institute/country	Horizontal resolution (lon × lat)
1	CESM2-FV2	National Center for Atmospheric Research (NCAR)/USA	2.5° × 1.9°
2	CESM2-WACCM-FV2	NCAR/USA	2.5° × 1.9°
3	FGOALS-g3	Chinese Academy of Sciences(CAS) /China	2° × ~ 5.18°
4	NESM3	Nanjing University of Information Science and Technology (NUIST)/China	1.875° × ~ 1.7°
5	SAM0-UNICON	Seoul National University (SNU)/Korea	1.25° × ~ 0.9°
6	CanESM5	Canadian Centre for Climate Modelling and Analysis (CCCma)/Canada	2.8125° × ~ 2.7673°
7	MCM-UA-1-0	Department of Geosciences, University of Arizona (UA)/USA	3.75° × ~ 2.22°
8	BCC-CSM2-MR	Beijing Climate Center (BCC)/China	1.125° × ~ 1.112°
9	CESM2-WACCM	NCAR/USA	1.25° × ~ 0.9°
10	MRI-ESM2-0	Meteorological Research Institute of Japan Meteorological Agency (MRI)/Japan	1.125° × ~ 1.11°
11	CESM2	NCAR/USA	1.25° × ~ 0.9°
12	MIROC6	Center for Climate System Research (CCSR), National Institute for Environmental Studies (NIES), Frontier Research Center for Global Change (FRCGC)/Japan	1.40625° × ~ 1.39°
13	BCC-ESM1	BCC/China	2.8125° × ~ 2.6763°
14	CAS-ESM2-0	CAS/China	1.40625° × ~ 1.41732°
15	GISS-E2-1-G	National Aeronautics Space Administration (NASA) Goddard Institute for Space Studies (GISS)/USA	2.5° × 2°
16	GISS-E2-1-H	NASA-GISS/USA	2.5° × 2°
17	CAMS-CSM1-0	Chinese Academy of Meteorological Sciences (CAMS)/China	1.125° × ~ 1.112°
18	MPI-ESM1-2-LR	Max Planck Institute for Meteorology (MPI-M)/Germany	1.875° × 1.85°
19	MPI-ESM-1-2-HAM	ETH Zurich, Switzerland; Max Planck Institut fur Meteorologie, Germany; Forschungszentrum Julich, Germany; University of Oxford, UK; Finnish Meteorological Institute, Finland; Leibniz Institute for Tropospheric Research, Germany; Center for Climate Systems Modeling (C2SM) at ETH Zurich, Switzerland (HAMMOZ-Consortium)	1.875° × 1.85°
20	MPI-ESM1-2-HR	MPI-M/Germany	0.9375° × ~ 0.9272°
21	AWI-ESM-1-1-LR	The Alfred Wegener Institute(AWI), Helmholtz Centre for Polar and Marine Research in Bremerhaven/Germany	1.875° × 1.85°
22	FIO-ESM-2-0	First Institute of Oceanography (FIO), Ministry of Natural Resources/China	1.25° × ~ 0.9°
23	AWI-CM-1-1-MR	AWI/ Germany	0.9375° × ~ 0.9272°
24	CMCC-CM2-SR5	Fondazione Centro Euro-Mediterraneo sui Cambiamenti Climatici (CMCC)/Italy	1.25° × ~ 0.9°
25	TaiESM1	Research Center for Environmental Changes, Academia Sinica (AS-RCEC), Taiwan/China	1.25° × ~ 0.9°
26	NorCPM1	Norwegian Climate Centre (NCC)/Norway	2.5° × ~ 1.89°
27	NorESM2-MM	NCC/Norway	1.25° × ~ 0.9°
28	NorESM2-LM	NCC/Norway	2.5° × ~ 1.89°
29	ACCESS-CM2	Australian Research Council Centre of Excellence for Climate System Science (ARCCSS)/Australia	1.875° × 1.25°
30	ACCESS-ESM1-5	Commonwealth Scientific and Industrial Research Organisation (CSIRO)/Australia	1.875° × 1.25°

$$SDR = \frac{SD(model)}{SD(OBS)},$$

$$(1) \quad Skill = \frac{(1 + R)^2}{(SDR + [1/SDR])^2}, \quad (2)$$

where R denotes the spatial correlation coefficient between the simulated (model) and observed (OBS) climatology, and

SDR is the ratio between the spatial standard deviation (SD) of the simulated climatology and that of the observations. The spatial pattern is considered when comparing model outputs with observations.

2.2.2 Moisture budget

To develop a deep understanding of the physical processes of future changes in summer precipitation, a diagnostic analysis of the vertically integrated moisture budget associated with precipitation from historical to future-warming simulations is performed. The moisture budget equation is as follows (Chou et al. 2009; Seager et al. 2010; Jiang et al. 2020; Moon and Ha 2020):

$$\frac{\partial \langle w \rangle}{\partial t} + \left\langle \nabla \cdot (q \vec{V}) \right\rangle = E - P + Res, \quad (3)$$

where w is precipitable water; $\vec{V}$ is the horizontal wind vector (zonal and meridional components); q is specific humidity; P is precipitation; E is evaporation; $\langle \rangle$ denotes vertical integration from the surface pressure to top-of-atmosphere pressure; and ∇ is the horizontal gradient operator. Although precipitable water w in Eq. (3) differs between historical and future periods, it is considered to be in an equilibrium state during both periods. Hence, the first term $\frac{\partial \langle w \rangle}{\partial t}$ on the left-hand side of Eq. (3) is omitted. Res is residual. The residual includes errors from computation, contributions from weather fluctuations, biases from the data assimilation system, and inconsistency between different datasets (Seo and Ok 2013). The residual term is omitted in this study.

The differences in the remaining terms between historical and future periods in Eq. (3) are responsible for future summer precipitation changes, as shown in Eqs. (4) and (5) Li et al. (2020), Zhao et al. (2019, 2020)

$$\Delta P = -\Delta \langle \nabla \cdot q \vec{V} \rangle + \Delta E, \quad (4)$$

$$\Delta P = -\Delta \langle \vec{V} \cdot \nabla q \rangle - \Delta \langle q \nabla \cdot \vec{V} \rangle + \Delta E, \quad (5)$$

where Δ denotes the difference between future and historical simulations (future minus historical). In Eq. (4), the first item $-\Delta \langle \nabla \cdot q \vec{V} \rangle$ on the right-hand side indicates changes in vertically integrated moisture flux convergence. On the right-hand side of Eq. (5), the first item $-\Delta \langle \vec{V} \cdot \nabla q \rangle$ denotes changes in moisture advection, referred to as “moisture pooling”; the second term $-\Delta \langle q \nabla \cdot \vec{V} \rangle$ is proportional to changes in moisture convergence, representing the total column integral of horizontal wind convergence, which is related to the lifting force. Therefore, moisture flux convergence in Eq. (4) is the sum of the moisture advection and moisture convergence terms in Eq. (5). The ΔE term on the right-hand side of Eqs. (4) and (5) indicates changes in evaporation. Precipitation changes may thus be caused by

changes in horizontal moisture advection, changes in moisture convergence related to vertical motion, and changes in surface evaporation.

3 Validation of CMIP6 models

3.1 Observed changes in summer precipitation in East Asia

Figure 1 shows the historical (1981–2014) geographical distribution of 850-hPa wind and precipitation climatologies in summer derived from observational datasets and the 30 CMIP6 model outputs. In the comparison, two precipitation datasets (GPCP and CMAP) and two 850-hPa wind-field datasets (NCEP-2 and ERA5) are employed. For the NCEP-2 and CMAP pair (Fig. 1a), precipitation maxima in the tropical regions are centered over the Indochina Peninsula, South China Sea (SCS), and western Pacific. Here, moist southwesterly winds from the SCS reach higher latitudes. A strong anticyclone dominates the western North Pacific; the intensified southerly flow along the southern flank of this anticyclone encounters the prevailing southwesterly flow, both of which transport abundant moisture. Previous studies have shown that the moist southerly flow may be well explained by the dynamic influence of the TP, in which forced topographic Rossby waves induce downstream southerlies. Specifically, the subtropical zonal wind impinging on the TP can result in cyclonic and anticyclonic circulation anomalies downstream, thereby inducing a substantial zonal geopotential height gradient, leading to southerly moisture transport to the EASM region (Son et al. 2019, 2020).

Accompanying the abundant moist southerly flow, a distinctive zonally elongated precipitation structure stretching from eastern China to Korea and Japan forms in the subtropics, known as “Changma” in Korea, “Meiyu” in China, and “Baiu” in Japan. In addition to these distinctive features in EA, a southeast–northwest tilted structure of the precipitation is observed in the northern part of EA, as reported by Qu (2017). In the other pair of observational datasets (GPCP and ERA5; Fig. 1b), although the precipitation intensity in GPCP is weaker than that in CMAP (Fig. 1a), the corresponding patterns of monsoon precipitation and circulation of these two observational datasets are similar. Overall, there is good agreement between the two observational datasets, with spatial correlation coefficients of 0.93 (NCEP-2 and ERA5) and 0.89 (GPCP and CMAP), respectively. The NCEP-2 and CMAP pair were selected as the observational reference in this study.

Next, the capabilities of the 30 individual CMIP6 models in simulating the EASM structure are presented in detail (Fig. 1). The southwesterly flow from the SCS agrees well

Pr&wind850

Fig. 1 Climatological summer (JJA) precipitation (shading; unit: mm d⁻¹) and 850-hPa wind field (vectors; unit: m s⁻¹) in EA from **a** CMAP/NCEP-2, **b** GPCP/ERA-5, and **c–z**, *I–VI* individual CMIP6 models during the historical period 1981–2014. Purple lines denote country boundaries

with the observations in all except two models (AWI-CM1-1-MR Fig. 1f and AWI-ESM1-1-LR Fig. 1g), although the low-level southwesterly flow in some models (ACCESS-CM2 Fig. 1c, ACCESS-ESM1-5 Fig. 1d, CAMS-CSM1-0 Fig. 1j, MPI-ESM1-2-HR Fig. 1z) is slightly stronger than the observations, contributing to more precipitation in the tropics. In contrast, a few models, such as CAS-ESM2-0 (Fig. 1i) and NorESM2-MM (Fig. 1v) underestimate the intensity and scope of tropical precipitation. Nearly all of the models can characterize the subtropical anticyclone over the western North Pacific and associated southerly moisture transport into East China, the Korean Peninsula, and Japan. Hence, the majority of models in CMIP6 can generate the subtropical rain-belt reasonably well. The spatial correlations between simulated precipitation in the 30 CMIP6 models and observed subtropical precipitation are compared (Figure omitted). The majority of models have a high correlation with observations when reproducing the subtropical precipitation, exceeding the 90% confidence level. In particular, the simulated precipitation in some models (ACCESS-ESM1-5 in Fig. 1d, CESM-FV2 in Fig. 1n, GISS-E2-1-G in Fig. 1t, CMCC-CM2-SR5 in Fig. 1q, MRI-ESM2-0 in Fig. 1l, NorCPM1 in Fig. 1iii, and TaiESM1 in Fig. 1vi) shows the significant correlation with observed subtropical precipitation, which is above the 95% confidence level. However, the western North Pacific anticyclone in a few models (NESM3 in Fig. 1ii and MCM-UA-1.0 in Fig. 1v) shifts northward, in association with a relatively weak precipitation front along the middle and lower reaches of the Yangtze River, the Korean Peninsula, and northern parts of Japan. Additionally, most models can reproduce the observed southeast–northwest tilted structures of the precipitation in northern EA. Compared with the observed precipitation over the EA landmass, almost all models overestimate precipitation, with excessive precipitation along the eastern periphery of the TP, as evidenced in previous studies (Yu et al. 2000; Wang et al. 2014). In summary, the majority of these 30 models can reliably reproduce the main EASM features, including abundant tropical precipitation, southwesterly monsoon flows, and the western North Pacific anticyclone.

3.2 Evaluation of CMIP6 model performance and selection of the best models

Despite the general agreement between the CMIP6 climate models and observations, substantial discrepancies are still

found in some areas of the study region. Many studies have reported a link between simulated monsoon circulation and precipitation (Sperber et al. 2013; Xin et al. 2020). To objectively quantify the fidelity of the 30 CMIP6 models in terms of representing the main EASM features, a Taylor diagram is shown in Fig. 2. In the Taylor diagram, the radial coordinate is the spatial SD of the model simulations divided by the SD of the observations. The spatial correlation coefficient between the observed and simulated fields is given by the angle from the y axis. Just over half of the CMIP6 models (16 of 30) simulate SDs lower than that of the observations, with SD < 0.6 in three models (BCC-CSM2-MR, NESM3, and MIROC6), whereas the SDs in some models (six of 30) are higher than that of the observations. Eight of the 30 models simulate SDs similar to the observations; in particular, the SDs in two models (CESM2-WACCM and SAM0-UNICON) are the same as the observations. The spatial correlation coefficients between the simulated and observed mean precipitation range from 0.56 to 0.84. The spatial correlation coefficients in seven models are above 0.8, with three models (CAMS-CSM1-0, NorCPM1, and MPI-ESM1-2-HR) having the highest correlation.

The spatial correlation coefficient and SD cannot fully cover all aspects of the EASM precipitation simulations. In our study, the evaluation of climate models mainly concerns the performance of CMIP6 models in simulating the spatial pattern of the observed EASM precipitation climatology in terms of amplitude and similarity. Accordingly, model performance is quantitatively assessed using the skill score metric defined in Eq. (2). This skill score method is suitable for quantifying the similarity and amplitude of the simulated spatial pattern of the EASM precipitation climatology with those of the observations, and has recently been widely used in evaluating model performance (Song and Zhou 2014a, b; Xin et al. 2020). The skill scores of the 30 CMIP6 models in reproducing the EASM precipitation pattern are shown in Fig. 3a, revealing a wide range from 0.53 to 0.81. Among these models, four models (CAMS-CSM1-0, GISS-E2-1-G, MRI-ESM2-0, and NorCPM1) perform well in reproducing the spatial pattern of summer precipitation climatology, with skill scores of 0.81, followed by TaiESM1 and ACCESS-ESM1-5 with relatively high scores of 0.78 and 0.77, respectively. In contrast, some models (MIROC6, NorESM2-MM, and NESM3) perform rather poorly, with skill scores below 0.6. Figure 3b shows the 25th–75th percentile ensemble quartile intervals of skill scores in precipitation distribution simulated by the 30 CMIP6 models. Models that perform well for the historical period are expected to have a better chance of producing more reliable future projections, although there is no guarantee. To find the best models suitable for future projections, the third quartile (75th percentile) is chosen here as the cut-off criterion. Eight of the 30 models meet the requirement, with skill scores exceeding 0.76 ($\geq$ the

Fig. 2 Taylor diagram of the spatial pattern of summer precipitation climatology in EA simulated by the 30 CMIP6 models, by the MMM of the 30 CMIP6 models (referred to as MMM30), the MMM of the eight best-performing CMIP6 models (referred to as MMM08), and the MMM of the five best-performing CMIP6 models (referred to as MMM05) during the historical period 1981–2014

75th percentile): CAMS-CSM1-0 ($S=0.81$); GISS-E2-1-G ($S=0.81$); MRI-ESM2-0 ($S=0.81$); NorCPM1 ($S=0.81$); TaiESM1 ($S=0.78$); ACCESS-ESM1-5 ($S=0.77$); CESM2-FV2 ($S=0.76$); and CESM2-WACCM ($S=0.76$). Based on the comparative analysis of the observed and simulated summer precipitation in EA, these eight models are selected to produce future projections.

Previous studies have documented that the MMM can alleviate biases in individual models, as well as outperform individual models (Jiang et al. 2005; Inoue and Ueda 2009). Thus, the MMM is employed to estimate the most likely features of the EASM precipitation in future projections. The geographical patterns of summer precipitation climatology, as shown by the MMM of the 30 models (referred to as MMM30) and of the eight high-skill models selected above (referred to as MMM08) are presented in Fig. 4b, c. In the tropical zone, although both MMM08 and MMM30 substantially underestimate summer precipitation, MMM08 performs slightly better than MMM30. For the subtropical region, MMM08 (Fig. 4c) shows better agreement with the observations (Fig. 4a) in reproducing the Meiyu/Changma/Baiu front characteristics relative to MMM30 (Fig. 4b). However, for the eastern coast of China, both MMM30 and MMM08 show a negative bias, producing weaker precipitation than the observed rain-belt extending from East China, which may be related to convective precipitation parameterization schemes and relatively inferior resolutions, as mentioned in a previous study (Sperber et al. 2013). In addition, both MMM08 and MMM30 (Fig. 4b, c) overestimate

precipitation across most of the EA continent, particularly on the eastern slope of the TP; this overestimate is a common feature in state-of-the-art coupled models (Lee et al. 2010; Lee and Wang 2012). Comparisons between MMM08 and MMM30 suggest that the performance of MMM08 (Fig. 4c) is better than that of MMM30 (Fig. 4b). Owing to resource constraints, future precipitation has to be projected with fewer models, as three (CESM2-FV2, NorCPM1, and TaiESM1) of the eight best-performing models have not participated in the Scenario Model Intercomparison Project (ScenarioMIP) of CMIP6. Therefore, the MMM of the other five best-performing models, CAMS-CSM1-0, GISS-E2-1-G, MRI-ESM2-0, ACCESS-ESM1-5, and CESM2-WACCM (hereafter referred to as MMM05) is adopted to obtain a more reliable future summer precipitation projection; MMM05 displays an excellent simulation skill (similar to MMM08) in reproducing the main EASM precipitation pattern (Fig. 4d). The Taylor diagram in Fig. 2 supports the result, showing that the positions of both MMM05 and MMM08 are closer to the reference point than any individual model or MMM30. Specifically, compared with a single model or MMM30, both MMM05 and MMM08 can (1) simulate a similar SD compared with that of the observations; (2) achieve higher spatial correlation coefficients with the observed EASM precipitation of nearly 0.9 (Fig. 2); and (3) achieve greater skill scores, exceeding 0.8 (Fig. 3). In contrast, MMM30 has a relatively weak correlation (~ 0.85) and simulates a lower SD of precipitation compared with that of the observations, achieving a lower skill score of only

Fig. 3 **a** Skill scores of the spatial pattern of summer precipitation climatology over EA in each model of CMIP6, and in MMM30, MMM08 and MMM05 during the historical period. **b** Box-and-whisker plot showing the minimum, 25th, 50th (median), 75th, and maximum percentiles of skill scores of summer precipitation climatology in EA simulated by the 30 models contributing to CMIP6 for the historical period

0.7. The advantage of MMM05 and MMM08 over MMM30 is understandable because poorly performing models are excluded in the calculations of MMM05 and MMM08.

The large differences in summer precipitation among EA sub-regions are worthy of mention (Fig. 4). These arise from modulation by land–sea thermal contrasts and pressure

gradients, the complex orographic structure of the EA land-mass, and various slowly varying components of the climate system such as global teleconnections, El Niño Southern Oscillation (ENSO), the Arctic Oscillation, tropical monsoons, Eurasian snow cover, TP thermal forcing, and sea surface temperature variability over adjacent oceans (Yu et al.

Fig. 4 Spatial pattern of summer precipitation climatology (shading; unit: mm d^{-1}) based on **a** the NCEP2/CMAP pair, **b** MMM30, **c** MMM08, and **d** MMM05. Boxes in **c** and **d** denote the three key

sub-regions: tropical region (10° – 20° N, 100° – 150° E); East China (25° – 34° N, 110° – 123° E); and the Japan and Korea sector (30° – 40° N, 125° – 150° E)

2000; Zhou et al. 2009; Ha et al. 2012; Wu et al. 2012). In addition, the horizontal distribution of air masses also plays a crucial role in the large differences in summer precipitation between EA sub-regions (Tomita et al. 2011; Seo et al. 2015). Usually, the mid-latitude monsoon rain-belt is influenced by the interaction of different air masses, but subtropical precipitation tends not to be. Therefore, a regional focus is beneficial in more reliably projecting the EASM precipitation. Wang et al. (2008) demonstrated that the first mode of EASM precipitation is closely linked to a meridional dipole pattern along the Yangtze River to Korea and Japan and in the Philippine Sea. The circulation and precipitation systems in these sub-regions are the main characteristics of the EASM. Wang et al. (2014) further divided EA into two parts: a tropical monsoon region from the SCS to the western Pacific, and a subtropical monsoon region covering the Meiyu/Changma/Baiu rain-belt. Ha et al. (2020) found different regional rates of increase of monsoon precipitation in response to global warming; specifically, the rate in the western North Pacific is less than that in northern EA. On this basis, our analysis of future projections of summer

precipitation in EA targets three key sub-regions: the tropical sector, including the Indochina Peninsula, the SCS, and the western Pacific (10° – 20° N, 100° – 150° E) (Li et al. 2010); East China, covering the middle and lower reaches of the Yangtze River Valley (25° – 34° N, 110° – 123° E); and the Japan and Korea sector (30° – 40° N, 125° – 150° E). In what follows, regional time series are calculated as area-weighted averages for each region.

4 Future changes in summer precipitation in the three sub-regions of East Asia

4.1 Temporal changes in future summer precipitation

There is high confidence of a warming trend during the twenty-first century (IPCC 2007, 2013; Zhan et al. 2020). Figure 5a shows the trend in summer mean near-surface air temperature series averaged across EA, as projected by MMM05 under SSP1-2.6, SSP2-4.5, SSP3-7.0, and

Fig. 5 **a** Time series of summer mean near-surface air temperature averaged in EA (unit: K; 11-year running mean) projected by MMM05 during the twenty-first century. **b–e** Time series of area-averaged summer precipitation anomalies (unit: mm d⁻¹) during the twenty-first century relative to the historical mean (1981–2014) in **b** EA, **c** East China, **d** Japan and Korea, and **e** the tropical region.

SSP5-8.5. Scenarios other than SSP1-2.6 project an increasing trend in the twenty-first century, and the trend is greater under higher-emission scenarios. The future summer precipitation changes across the sub-regions of EA under the four scenarios show notable differences in time and region (Fig. 5b–f). In Fig. 5b, summer precipitation averaged across the whole of EA under all the scenarios shows an upward trend throughout the twenty-first century, with responses yielding increases of 7.3%, 9.1%, 9.1%, and 14.5% for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively. During the near- to mid-term, a moderate increase can be seen under

Plotted projections are by MMM05 for SSP1-2.6 (black), SSP2-4.5 (blue), SSP3-7.0 (green), and SSP5-8.5 (red). **f** Percentage changes (unit: %) of 10-year mean summer precipitation from 2020 to 2095 under SSP5-8.5 projected by MMM05 relative to the historical mean in East China (black), Japan and Korea (green), the tropical region (red), and EA (blue)

the four scenarios. After the 2060s, a prominent increase is identified under SSP5-8.5, compared with relatively slow increases under the other scenarios. That is, there are evident distinctions in precipitation changes among the four scenarios in the long-term, with SSP5-8.5 giving a more pronounced upward trend. This precipitation trend across the whole of EA has already been recognized under a warming climate; however, this study reveals the trends in the three sub-regions.

As seen in Fig. 5c, summer precipitation in East China under all the scenarios is expected to increase by the end of

this century, at rates of 9.3%, 11.1%, 9.3%, and 14.8% under SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively. Differences between the four scenarios are indistinguishable until the 2090s, followed by an exponential increase under SSP5-8.5. In Japan and Korea, precipitation is projected to increase throughout this century under SSP2-4.5, SSP3-7.0, and SSP5-8.5. The response rates by 2100 are 5.4%, 8.9%, 7.1%, and 10.7% for SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively, (Fig. 5d). In the tropical region, the response rates correspond to 4.3%, 7.4%, 7.4%, and 14.9% in the twenty-first century under SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5, with precipitation under SSP5-8.5 always the greatest (Fig. 5e). Moreover, the temporal changes of future summer precipitation projected by MMM30 under the four scenarios show similar trends to those of MMM05 (Figure omitted). Some differences in future summer precipitation changes exist between MMM30 and MMM05, which are mainly due to uncertainties in the simulations over the EASM region (Kitoh et al. 2013; He and Zhou 2015).

Nonlinear hydrological responses to warming vary among the different scenarios, as discussed by Wu et al. (2010). A comparison of summer precipitation trends under warming across the different sub-regions reveals that precipitation across all sub-regions in the long-term under SSP5-8.5 is generally far greater than that under the other scenarios. Hereafter, we focus only on the results of SSP5-8.5. The 10-year mean summer precipitation anomaly relative to the historical mean under SSP5-8.5 is shown in Fig. 5f. All of the most prominent fluctuations in EA and the three sub-regions occur after the 2070s, which demonstrates that the magnitudes of precipitation variations after the 2070s are greater than those in any preceding decades, and

also indicates the classification into the near- to mid-term (2020–2069) and long-term (2070–2095) is consistent with the temporal characteristics of future summer precipitation changes in the sub-regions of EA. Results further suggest that the long-term precipitation has stronger sensitivity to warming than that of the near- to mid-term precipitation. This feature may be a unique regional response in EA, as it has not been found in other areas (Sun and Ding 2010; Woo et al. 2018).

4.2 Spatial changes in future summer precipitation

Another crucial question is what the dominant summer precipitation patterns in the sub-regions of EA will look like under future warming. Figure 6 shows the geographical distribution of projected summer precipitation anomalies during the near- to mid-term and long-term under SSP5-8.5 relative to the historical mean. The near- to mid-term precipitation in EA, the tropics, East China, Japan and Korea, shows a moderate increase (by about 7.3%, 7.4%, 3.7%, and 0.18%, respectively; Fig. 6a). After the 2070s, summer precipitation intensifies markedly across all three sub-regions. In East China, the region of increased precipitation expands slightly northward with intensified precipitation relative to the near- to mid-term. The large region of Japan and Korea is expected to experience more positive precipitation anomalies in the long-term (Fig. 6b). The long-term precipitation in the whole of the EA region, the tropics, East China, Korea and Japan has response rates of 14.5%, 14.9%, 14.8%, and 10.7%, respectively. In brief, future summer precipitation in the three key EA sub-regions where the wet pattern dominates in the historical period is projected to enhance

Fig. 6 Percentage changes in future summer precipitation (unit: %) projected by MMM05 under SSP5-8.5 in **a** the near- to mid-term, and **b** the long-term relative to the historical mean (1981–2014). Dots

indicate areas where at least 66% of the selected models agree on the sign of the MMM changes. Boxes denote the three key sub-regions

more significantly during the long-term than in the near- to mid-term.

5 Mechanisms for changes in future summer precipitation: the moisture budget perspective

The remarkable difference in future summer precipitation changes between the near- to mid-term and long-term motivates us to investigate the potential hydrological processes. To quantitatively analyze the contributions of the various factors accounting for such changes, the moisture budget equation (Eq. (5)) is employed. The future summer precipitation changes under warming feature an increasing trend in the three sub-regions of EA, and are dominated by increasing moisture convergence, while the evaporation and moisture advection terms contribute weakly to the increasing precipitation trend (Fig. 7). The moisture convergence is expected to play a primary role in future summer precipitation increases across the three sub-regions of EA.

As atmospheric moisture and circulation changes can influence moisture convergence, it is necessary to further

analyze their contributions. The changes in moisture convergence can be decomposed into three terms, as follows (Hsu et al. 2012):

$$-\Delta\langle q \cdot D \rangle = -\langle q_{pd} \cdot \Delta D \rangle - \langle \Delta q \cdot D_{pd} \rangle - \langle \Delta q \cdot \Delta D \rangle, \quad (6)$$

where $\langle \rangle$ refers to the mass integration from surface pressure (1000 hPa) to the top-of-atmosphere pressure at 100 hPa; Δ denotes the discrepancy between future and historical simulations (future minus historical simulation); D (representing $\nabla \cdot \vec{V}$) denotes divergence; and ∇ represents the horizontal gradient operator. The subscript “pd” means the historical simulation. The first term on the right-hand side of Eq. (6) represents circulation changes; thus, it reflects the dynamic component of future summer precipitation changes. The second term on the right-hand side of Eq. (6) is associated with changes in specific humidity (moisture content) and may be regarded as the thermodynamic component. The third term on the right-hand side of Eq. (6) represents the nonlinearity resulting from the combined effect of both moisture and circulation changes and can be assumed to be small. The moisture budget decomposition can also

Fig. 7 Changes in future summer precipitation ΔP and relative contributions of moisture advection $-\Delta \langle \vec{V} \cdot \nabla q \rangle$, moisture convergence $-\Delta \langle q \nabla \cdot \vec{V} \rangle$, and surface evaporation ΔE to future summer precip-

itation changes ΔP in **a** EA, **b** East China, **c** Japan and Korea region, and **d** the tropical region during the near- to mid-term and long-term. Unit is mm d^{-1}

be simplified as $\Delta P \sim -(\Delta q \cdot \omega + q \cdot \Delta \omega)$, where ω is the vertical velocity (Huang and Xie 2015a; Zhou et al. 2018).

The thermodynamic component makes the dominant positive contribution to the pronounced precipitation increase in the three sub-regions of EA, throughout the twenty-first century (Fig. 8). In contrast, the dynamic component during the near- to mid-term offsets the precipitation increases in subtropical regions including East China, Japan and Korea (Fig. 8b, c), but favors tropical precipitation increase (Fig. 8d). Over the long-term, the dynamic component still contributes negatively to precipitation in Japan and Korea, and makes a positive contribution to precipitation increase in the tropics (Fig. 8c, d), but exerts a much weaker effect in East China (Fig. 8b). Overall, the thermodynamic component is a contributor to future summer precipitation increases in the three sub-regions of EA, whereas the strength of the dynamic component varies between different sub-regions. As for the geographical distribution, both the near- to mid-term and long-term show that the thermodynamic component is almost positive in the three sub-regions of EA, with the magnitude in the long-term (Fig. 9c) being much greater than that in the near- to mid-term (Fig. 9a). However, the

sign of the dynamic component varies notably in the three sub-regions (Fig. 9b, d).

The heterogeneous distribution of the dynamic component motivates us to investigate whether it is offset by inter-model uncertainty. We analyzed the uncertainty of the thermodynamic and dynamic components to assess this possibility. The description of model uncertainty is introduced as follows. The inter-model SD (σ) of changes is employed to quantify the inter-model spread. To examine the credibility of changes projected by the MMM, the signal-to-noise ratio (SNR), defined as the ratio of changes projected by MMM (implying signal) to inter-model spread (σ , implying noise or uncertainty) is used in this study (Long et al. 2016; Zhou et al. 2018; Liu et al. 2020). The model agreement is the percentage of changes projected by the individual models that agree with the sign of changes projected by MMM. A threshold of 66% is employed herein, implying the “likely” changes (Mastrandrea et al. 2010; Power et al. 2012; Chen et al. 2014).

$$SNR = \frac{\Delta}{\sigma}, \quad (7)$$

Fig. 8 Relative contributions of the dynamic component— $\langle q_{pd} \cdot \Delta D \rangle$, thermodynamic component— $\langle \Delta q \cdot D_{pd} \rangle$, and the nonlinear term arising from both circulation and moisture

changes— $\langle \Delta q \cdot \Delta D \rangle$ to moisture convergence— $\Delta \langle q \nabla \cdot \vec{V} \rangle$ in **a** EA, **b** East China, **c** Japan and Korea region, and **d** the tropical region during the near- to mid-term and long-term. Unit is mm d^{-1}

Fig. 9 Spatial distribution of the **a, c** thermodynamic and **b, d** dynamic components of future summer precipitation changes projected by MMM05 in EA during the near- to mid-term and long-term. Unit is mm d^{-1} . Boxes denote the three key sub-regions

where Δ indicates the changes projected by MMM, and σ indicates the inter-mode SD of changes. Large (smaller) SNR denotes relatively little (great) inter-model spread compared to the MMM changes and corresponds to high (low) reliability in model projections.

Figure 10 displays the inter-model spread of thermodynamic and dynamic components, and the corresponding SNR. In general, the inter-model spread is greater at lower latitudes than at mid to high latitudes. The inter-model spread in the thermodynamic component during both the near- to mid-term and long-term (Fig. 10a, b) is lower than 0.5 mm/day in most locations and is much smaller than that of the dynamic component (which exceeds 1.0 mm/day in most of the sub-regions; Fig. 10c, d). The centers of large spread in the dynamic component are mainly located in the southern coast of China and tropical region, exceeding

1.5 mm/day (Fig. 10c, d). The SNR of the thermodynamic component has high values above 1.5 over most of the sub-regions during the near- to mid-term and long-term (Fig. 10e, f). These large-value regions are generally collocated with large precipitation change (signal) regions, indicating that the future precipitation changes and related thermodynamic component in the three sub-regions are quite robust. In contrast, the uncertainty in future summer precipitation changes is primarily attributed to the large inter-model spread in the dynamic component, with the absolute value of SNR less than 0.5 across the whole of EA (Fig. 10g, h). Additionally, the thermodynamic component has > 66% model agreement across the whole of EA, including the three sub-regions (Fig. 10e, f), whereas the areas with the dynamic component exceeding the 66% model agreement only cover certain regions to the north of East China, Korea

Fig. 10 a–d Inter-model spread (noise) and e–h SNR of the thermodynamic and dynamic components of future summer precipitation changes during the near- to mid-term and long-term, respectively.

Dots indicate regions where >66% of models agree on the sign of the MMM changes. Boxes denote the three key sub-regions

and Japan (Fig. 10g, h). The above results indicate that the contribution from the dynamic component to future summer precipitation changes seems to be relatively weak, which may be canceled out by the large difference in dynamic components between various models.

The robust contribution of the thermodynamic component in increasing future summer precipitation over the three sub-regions of EA is found to dominate that of the dynamic component, which shows large uncertainty. Detailed analysis is further performed to investigate the possible causes underlying the changes in the thermodynamic and dynamic components (Fig. 11). The spatial pattern in the thermodynamic component (Fig. 9a, c) is dominated by climatological vertical velocity (ω ; Fig. 11g) and is mediated by a spatial field of increasing moisture (Δq ; Fig. 11e, f), which has > 66% model agreement in all three sub-regions. Both the pattern and magnitude of the dynamic component can be attributed to the changes in vertical velocity, but with > 66% model agreement in only a few regions ($\Delta\omega$; Fig. 11h, i).

Through the comprehensive analysis of future changes in moisture content, moisture flux convergence, and the pattern of precipitation minus evaporation under the warming climate, the pattern of “wet-getting-wetter” is clear: during the near- to mid-term and long-term of the twenty-first century, the specific humidity increases (Fig. 11e, f) uniformly across EA under a warming climate (Fig. 11b, c). The thermodynamic component, mediated by increased humidity, thus exhibits significant increases in the three sub-regions of EA (Fig. 9a, c). The enhanced thermodynamic component then contributes to the remarkable moisture flux convergence changes. Therefore, the existing pattern of $P-E$ (Eq. (4); Fig. 11n, o) is expected to strengthen via the enhancement of anomalous moisture flux convergence (Fig. 11k, l) in the three sub-regions of EA, where the existing pattern of $P-E$ is wet in the historical period (Fig. 11m).

6 Conclusions and discussions

Using the newly released CMIP6 simulations from 30 models, historical observations, and reanalysis, this study has examined the model skill in simulating the spatial pattern of observed summer precipitation climatology in three sub-regions of EA. A subset of the best-performing models was then used to investigate the future changes in summer precipitation under four emission scenarios (SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5), revealing possible changes in summer precipitation and associated processes underlying the projected changes.

The main findings are summarized as follows.

1. By evaluating the ability of the 30 CMIP6 models to reproduce the main EASM precipitation features, the five best-performing models are selected. These models (CAM5-CSM1-0, GISS-E2-1-G, MRI-ESM2-0, ACCESS-ESM1-5, and CESM2-WACCM) demonstrate skill scores exceeding the 75th percentile of the models participating in the CMIP6 ScenarioMIP. Three key sub-regions of EA are chosen for further analysis; i.e., the tropical sector, East China, and the Japan and Korea region, where the maximum centers of observed summer precipitation climatology in the historical period are located.
2. A comparison of projected summer precipitation trends in the three key sub-regions under the four scenarios show that the precipitation increases under SSP5-8.5 are slightly greater than those under the other scenarios in the long-term. Precipitations under SSP5-8.5 across the three sub-regions are expected to increase more significantly in the long-term, when compared with increases in the near- to mid-term. The response rates of projected precipitation in the tropics, East China, Japan and Korea are only about 7.4%, 3.7%, and 0.18%, respectively in the near- to mid-term, but show the intense increase by 14.9%, 14.8% and 10.7% in the long term.
3. According to the moisture budget, the significant increase in precipitation with different response rates in the three sub-regions of EA is primarily attributed to the thermodynamic component. Detailed analysis indicates that the pattern of the thermodynamic component is dominated by the climatological vertical velocity, mediated by a spatial increase in moisture. In contrast, the contribution of the dynamic component (related to circulation changes) is weak (Fig. 12). Both the pattern and magnitude of the dynamic component are mainly governed by changes in vertical velocity, although the uncertainty is large.

As mentioned above, the dynamic effects on projected monsoon precipitation changes in EA differ between various studies: some studies reported a positive contribution from the dynamic component (Wang et al. 2014; Li et al. 2019a), whereas others identified a negative one (Li et al. 2014; Lee et al. 2017). Our results have revealed that the contribution from the dynamic component in the three sub-regions may be canceled out by the large spread between different models, which can explain the difference in dynamic effects documented in previous studies. More importantly,

Fig. 11 Historical mean (1981–2014) of **a** near-surface air temperature (unit: K), **d** vertically integrated specific humidity (unit: kg m^{-2}), **g** vertical velocity at 500 hPa (unit: Pa s^{-1}), **j** vertically integrated moisture flux convergence (unit: $10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$), and **m** $P-E$ (unit: mm d^{-1}) in summer. Future changes of **b–c** near-surface air temperature (unit: K), and **n–o** $P-E$ (unit: mm d^{-1}) in summer projected by MMM05 under SSP5-8.5 in the near- to mid-term and the long-term. Percentage changes of **e–f** future vertically integrated specific humidity (unit: %), **h–i** vertical velocity at 500 hPa (unit: %) and **k–l** vertically integrated moisture flux convergence (unit: %) in summer projected by MMM05 under SSP5-8.5 in the near- to mid-term and the long-term. Dots indicate areas where at least 66% of the selected models agree on the sign of the MMM changes. Boxes denote the three key sub-regions

our study has focused on quantifying the contribution of thermodynamic and dynamic components to projected summer precipitation changes, and on analyzing the possible causes driving the changes in the two components using the simulations of the five selected models. As for the sources and correction of the large spread between the individual models, it is a topic of great interest in the climate community and warrants further focused investigation.

Some studies have utilized various approaches to reduce the inter-model spread in model projections. For example, to enhance the credibility of projections of tropical Pacific SST warming, a novel method of spatial emergent constraints with the ensemble pattern regression method has

been employed to reduce the common change bias (Huang and Ying 2015b). Similarly, the simulated bias from the EASM intensity and associated precipitation has been corrected to give a more reliable projection under global warming of 1.5 °C or 2 °C (Liu et al. 2020). Therefore, a bias-corrected method such as emergent constraints can be used to reduce the large spread in the dynamic component in future analysis.

Circulation changes, especially those associated with the summer monsoon, are critical to the dynamic component. A self-organizing map has been used to divide the annual circulation cycle into eight patterns based on the 850-hPa wind fields and investigate the onset dates of the monsoon stages and associated circulation patterns (Dai et al. 2021). Ha et al. (2020) projected an earlier onset and later retreat of the EASM, leading to an increase in the length of the summer monsoon in EA. Further study is required to address the influence of annual cycle changes on projected monsoon precipitation.

Finally, it is important to keep in mind that this study focused on mean changes at the monthly time scale. Our conclusions thus rely strongly on the assumption of negligible non-linear interaction of shorter-term co-variations of the variables involved. Therefore, the next step may be to use daily or even hourly datasets to verify the results.

Fig. 12 A schematic diagram summarizing the main mechanisms responsible for the projected increasing summer precipitation in the three key sub-regions of EA from the perspective of moisture budget analysis

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