



Distinct response of Asian summer monsoon circulation and precipitation to orbital forcing during six Heinrich events

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ABSTRACT

Climatic fingerprint of Heinrich (H) events was characterized by widespread megadroughts over the Asian summer monsoon (ASM) region accompanied by systemic weakening of the ASM. However, recent studies of hydroclimate proxies suggest huge spatial discrepancies in precipitation over the ASM region during some H events, characterized by increased precipitation in the Yangtze River Valley contrasting with the prevalent megadroughts across the whole ASM region. The mechanism responsible for the spatial discrepancies in precipitation and the relationship between local precipitation and the ASM intensity remain elusive. In this study, we investigate the response of the ASM circulation and precipitation to orbital forcing during six H events based on simulations with a coupled atmosphere-ocean general circulation model. The results show that changes in insolation alone can induce spatial discrepancies in precipitation over the ASM region during the H events. During the H1, 3, 4, 5, 6 events, the amplification of the land-sea pressure contrast in response to a positive interhemispheric insolation gradient (30°N–30°S) during boreal summer intensifies moisture transport from the adjacent oceans to the ASM region. The ensuing moisture divergence, combined with anomalous downdrafts, results in decreased precipitation in the South Asian Summer Monsoon (SASM) region, but converse scenario for the East Asian Summer Monsoon (EASM) region. During the H2 event, the increased precipitation across the Yangtze River Valley sharply contrasts the widespread drought over the ASM region, attributing to an anti-cyclone anomaly over the subtropical Western North Pacific and a cyclone anomaly over Japan and Korea. Moisture budget analysis shows that the dynamic effect, especially the vertical term, rather than the thermodynamic effect, is the dominant control of precipitation changes over the ASM region. Our results also suggest that despite the synchronous variation in the strength of the EASM and SASM in response to orbital forcing, the EASM should not be regarded as an eastward and northward extension of the SASM. Furthermore, our model simulates a weak correlation between the monsoon intensity and precipitation in the SASM region in response to orbital forcing, calling for caution in employing precipitation to reconstruct SASM intensity on orbital time scale.

1. Introduction

The most-known abrupt climate changes during the last glacial period are the Dansgaard-Oeschger (DO) events and the Heinrich (H) events. The DO events are characterized by a decadal scale surface air temperature (SAT) increase of 5–16 °C in Greenland which is followed by a gradual cooling over a few centuries to a millennium and then an

abrupt drop to a cold stadial (Dansgaard et al., 1993). Notably, six of these extremely cold stadials—known as H events—were characterized by the occurrence of extensive ice-rafted debris in North Atlantic sediments, indicating an enhanced discharge of icebergs from the circum-North Atlantic continental ice sheets (Fig. 1f-black line) (Heinrich, 1988; Bond et al., 1992; Hemming et al., 2004). A massive iceberg discharge reduces Atlantic Meridional Overturning Circulation (AMOC)

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(Fig. 1f–red line, McManus et al., 2004; Böhm et al., 2015; Henry et al., 2016) and associated northward ocean heat transport, generating a strong widespread cooling in the Northern Hemisphere (Fig. 1a–purple line, North Greenland Ice Core Project Members, 2004) but warming in

Antarctica (Fig. 1b–cyan line, WAIS Divide Project Members, 2015). Meanwhile, a southward displacement of the Intertropical Convergence Zone (ITCZ) is observed (Fig. 1c–green line, Peterson et al., 2000), which is accompanied by a weakening of the Asian summer monsoon (Fig. 1d,

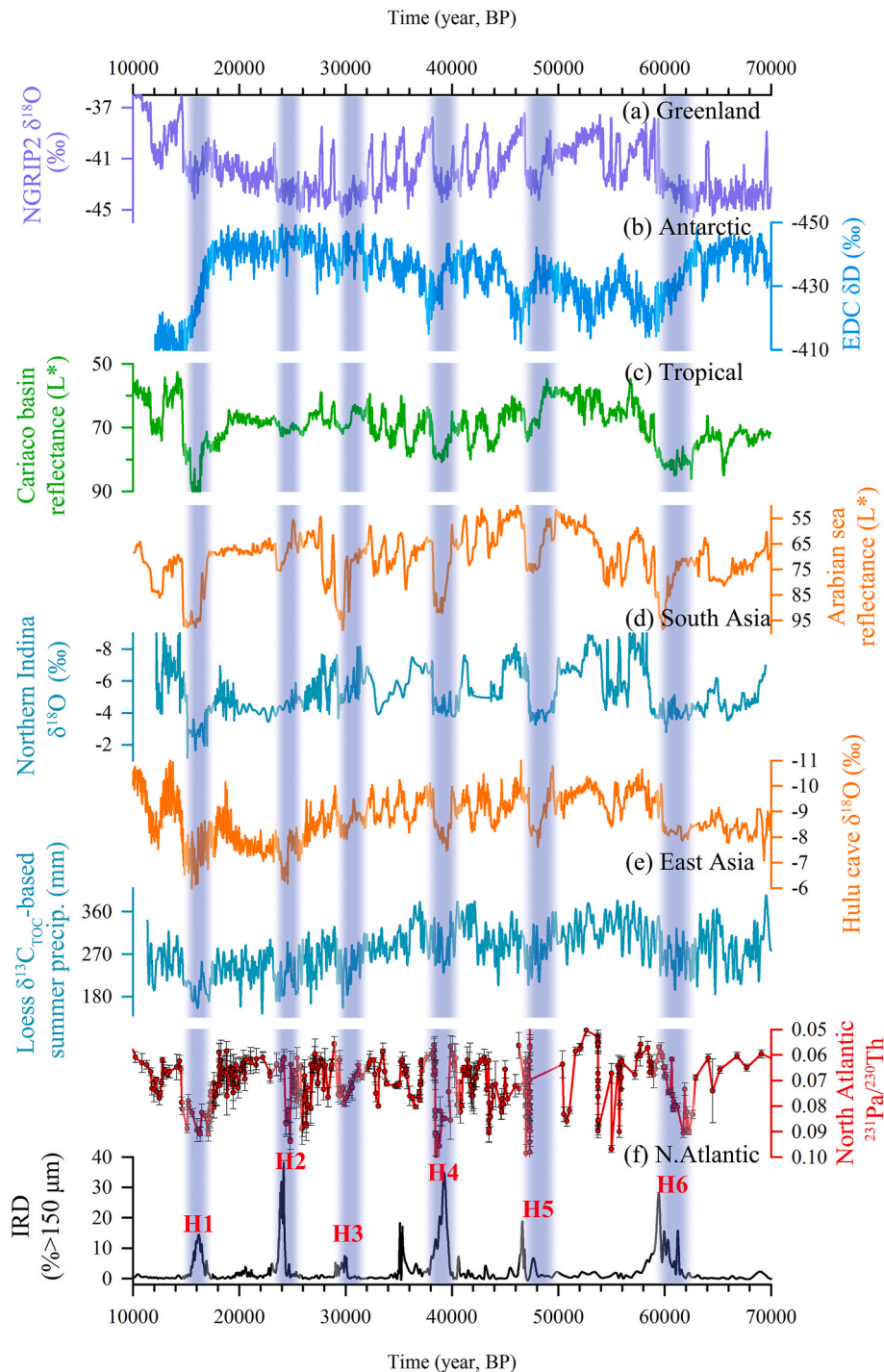


Fig. 1. The evolution of Asian summer monsoon and polar climate during the last glacial. Dark blue bars indicate the H1 (15.6–18.0 ka), H2 (24.3–26.5 ka), H3 (31.3–32.7 ka), H4 (38.3–40.2 ka), H5 (47.0–50.0 ka) and H6 (60.1–63.2 ka), respectively (Sanchez Goñi and Harrison, 2010). (a) Greenland: NGRIP $\delta^{18}\text{O}$ record (purple, NGRIP Community Members, 2004). (b) Antarctic: EDC δD record (cyan, WAIS Divide Project Members, 2015). (c) Tropical: Cariaco basin sediment total reflectance indicates the migration of the ITCZ (green, Deplazes et al., 2013). (d) South Asia: Arabian Sea sediment total reflectance as a proxy for SASM strength (Orange, Deplazes et al., 2013) and SASM rainfall record inferred from speleothem $\delta^{18}\text{O}$ in Bittou cave (Dark cyan, Kathayat et al., 2016). (e) East Asia: Hulu-Sanbao cave composite speleothem $\delta^{18}\text{O}$ records as a proxy for EASM strength (Orange, Cheng et al., 2016) and EASM rainfall records inferred from organic $\delta^{13}\text{C}$ in Chinese loess plateau (Dark cyan, Rao et al., 2013). (f) North Atlantic: observed $^{231}\text{Pa}/^{230}\text{Th}$ ratio from the deep North Atlantic as a proxy for AMOC (red line with error bar, McManus et al., 2004; Böhm et al., 2015; Henry et al., 2016), and the percentage of lithic grains in the $>150\ \mu\text{m}$ fraction in sediment core MD95-2040 as a symbolize of H event (black, De Abreu et al., 2003).

e-orange line, Wang et al., 2001a,b; Burns et al., 2003; Yuan et al., 2004; Cai et al., 2015; Cheng et al., 2016) and widespread dryness across the Afro-Asian monsoon region (Fig. 1d, e-dark cyan line, Stager et al., 2011).

In contrast to the prevalent megadroughts across the ASM region such as North China (Fig. 2a-green and dark cyan line) and South Asia (Fig. 2c-green line), a recently published moisture proxy, utilizing trace-element ratios (Sr/Ca, Mg/Ca, Ba/Ca) and speleothem $\delta^{13}\text{C}$ from the Haozhu Cave in the Yangtze River Valley, indicates a wet condition over the Yangtze River Valley region during the H1 event (Fig. 2b-green and dark cyan line) (Zhang et al., 2018). This is further confirmed by model simulation (He et al., 2021, Fig. 2a, b, c-red and green line). In addition, wetter climate in the Yangtze River Valley region has also been found in the H4 (Wu et al., 2023) and H5 (Liu et al., 2022) events. This “Yangtze River Valley flood and South, North China drought” triple pattern is consistent with the modern interannual rainfall variability in EASM region (Ding et al., 2008). Therefore, paleoclimate proxies and simulations indicate huge spatial discrepancies in regional precipitation in the ASM domain despite the prevailing drought condition during the H events. The most common explanation for the spatial discrepancies is that the massive freshwater release from the melting icebergs initially weakens the EASM strength associated with a seasonal delay in the northward migrating westerlies, a lengthening of the Meiyu rainfall, and a shortened midsummer stage (Zhang et al., 2018). However, the surface ocean temperature and ice-rafted debris (IRD) records from the north-east Atlantic indicate that the surface ocean cooling occurred hundreds to thousands of years before the IRD arrived in the North Atlantic, suggesting that the iceberg discharge (or freshwater) may be a consequence of stadial conditions rather than the cause (Barker et al., 2015). Thus, the mechanism of the spatial discrepancies in regional precipitation during the H events remains elusive. Considering that insolation is an important known external forcing of the climate system, it is worthwhile to test whether insolation change alone could contribute to the spatial discrepancies in precipitation over the ASM region during the H events.

The ASM, including two interactive sub-components: the EASM and the SASM, is one of the most important components of the global monsoon system (Trenberth et al., 2000; Wang et al., 2009) and characterized by a seasonal reversal of the prevailing winds accompanied by the contrast of rainy summer and dry winter (Webster et al., 1998). On orbital time scale, Kutzbach (1981) proposed that ASM varies in response directly to the intensity of North Hemisphere summer insolation (NHSI) caused by changes of Earth's orbital parameters. This was confirmed by the accurate dating and continental-scale coherence of speleothem $\delta^{18}\text{O}$ records in the ASM region, indicating that the variations of the ASM strength have a dominant ~ 21 -kyr precession cycle (Wang et al., 2001a,b; Burns et al., 2003; Fleitmann et al., 2003; Yuan et al., 2004; Cai et al., 2015; Kathayat et al., 2016; Cheng et al., 2016). In contrast, marine proxy records from the Arabian Sea, such as upwelling-driven biological productivity, foraminifera assemblage, particle size, and barium flux, which are interpreted as the SASM strength, indicate a ~ 8 -kyr time lag of SASM in response to the NHSI variations on the precession time scale (Clemens et al., 1991; Clemens and Prell, 2003; Caley et al., 2011). This approximately 8-kyr time lag is proposed to be caused by global ice volume (~ 5 kyr) and the cross-equatorial transport of latent heat from the South Hemisphere (~ 11 kyr), which is forced by the South Hemisphere local summer insolation (Clemens et al., 1991; Clemens and Prell, 2007). In addition, this “sea-land precession phase paradox” (Cheng et al., 2021) can also be found on millennial time scale. For example, accurately-dated speleothem $\delta^{18}\text{O}$ records indicate that during the onset of the H1 event, the EASM weakening occurred (Fig. 2a and b, red triangle with dash line) several hundred years before the SASM weakening (Fig. 2c, pink triangle with dash line) as inferred from the Arabian Sea sediment total reflectance. Therefore, the contradiction between the speleothem $\delta^{18}\text{O}$ and Arabian marine records raises a question of whether EASM is in/out of

phase with the SASM under orbital forcing on millennial (such as the H events) to orbital time scale.

This study aims to have a comprehensive understanding of the response of the ASM circulation and precipitation to orbital forcing during six H events (H1 to H6), based on six snapshot simulations performed with the HadCM3 model. In Section 2, the model, experiment setup and analysis method are introduced. In Section 3, the different effects of orbital forcing on the ASM circulation and precipitation during the H events are investigated. The decomposition of the physical processes controlling precipitation and associated large-scale circulation change is performed, and then the linkage of the large-scale circulation (such as the migration of the ITCZ, the ASM strength) and local summer precipitation is analyzed. A summary and conclusion are provided in Section 4.

2. Method

2.1. Model description and experiment setup

A fully coupled atmosphere–ocean general circulation model HadCM3 (Pope et al., 2000; Gordon et al., 2000) from the U.K. Met Office Hadley Center for Climate Prediction and Research, is used to investigate the different response of Asian monsoon circulation and precipitation to orbital forcing during six Heinrich events. The horizontal resolution of the atmospheric component is 3.75° longitude $\times$ 2.5° latitude with 19 unevenly spaced vertical levels and the ocean component has a horizontal resolution of 1.25° longitude $\times$ 1.25° latitude with 20 unevenly spaced vertical levels. Vertical levels are distributed to allow greater resolution closer to the surface in atmosphere and ocean model (Gordon et al., 2000). The land surface scheme is MOSES2.1 (Met Office Surface Exchange Scheme; Cox et al., 1999; Essery et al., 2003) which principally deals with the hydrology of the canopy to the subsurface and the surface energy balance, including subsurface thermodynamics. It also contains a simple thermodynamic sea-ice model based on Cattle and Crossley (1995) with parameterizations for ice drift and concentration.

The performance of HadCM3 in simulating the present climate over the ASM region is assessed in appendix model performance and Fig. A1. It shows that HadCM3 captures well the main features of observed present climatology over the ASM region. As mentioned in Section 1, the H events are mostly associated with massive iceberg discharge and freshwater flux. However, when investigate regional climate response during the H events, other forcing factors should also be considered, in particular orbital forcing which is quite different between the H events. As stressed in Section 1, this study aims to investigate the effect of orbital forcing on the ASM during the six H events, so only orbital forcing is considered in our simulations. All other configurations such as GHG (CO_2 : 280 ppmv, CH_4 : 760 ppbv, and N_2O : 270 ppbv), ice sheets and vegetation are fixed as the PI condition and no freshwater flux is added in the model. This allows to investigate the effect of orbital forcing alone. We conducted six time-slice experiments with time-dependent orbital forcing for the last glacial H events as shown in Table 1. The duration of each H event occupied 2–3 ka, and their top and base ages are defined according to Sanchez Goñi and Harrison, (2010)(Table 1). The mean values of the orbital parameters averaged over each H event were used in our simulations. All simulations are integrated for 500 model years, so the atmosphere and upper ocean reach an equilibrium state, and the climatology of the last 60 years is considered in the analysis.

2.2. Moisture budget analysis

To examine the physical processes governing the Asian summer monsoon precipitation changes in response to orbital forcing, neglecting the tendency term in the steady state, the time-mean column-integrated moisture budget equation can be written as (Seager et al., 2010; Chou

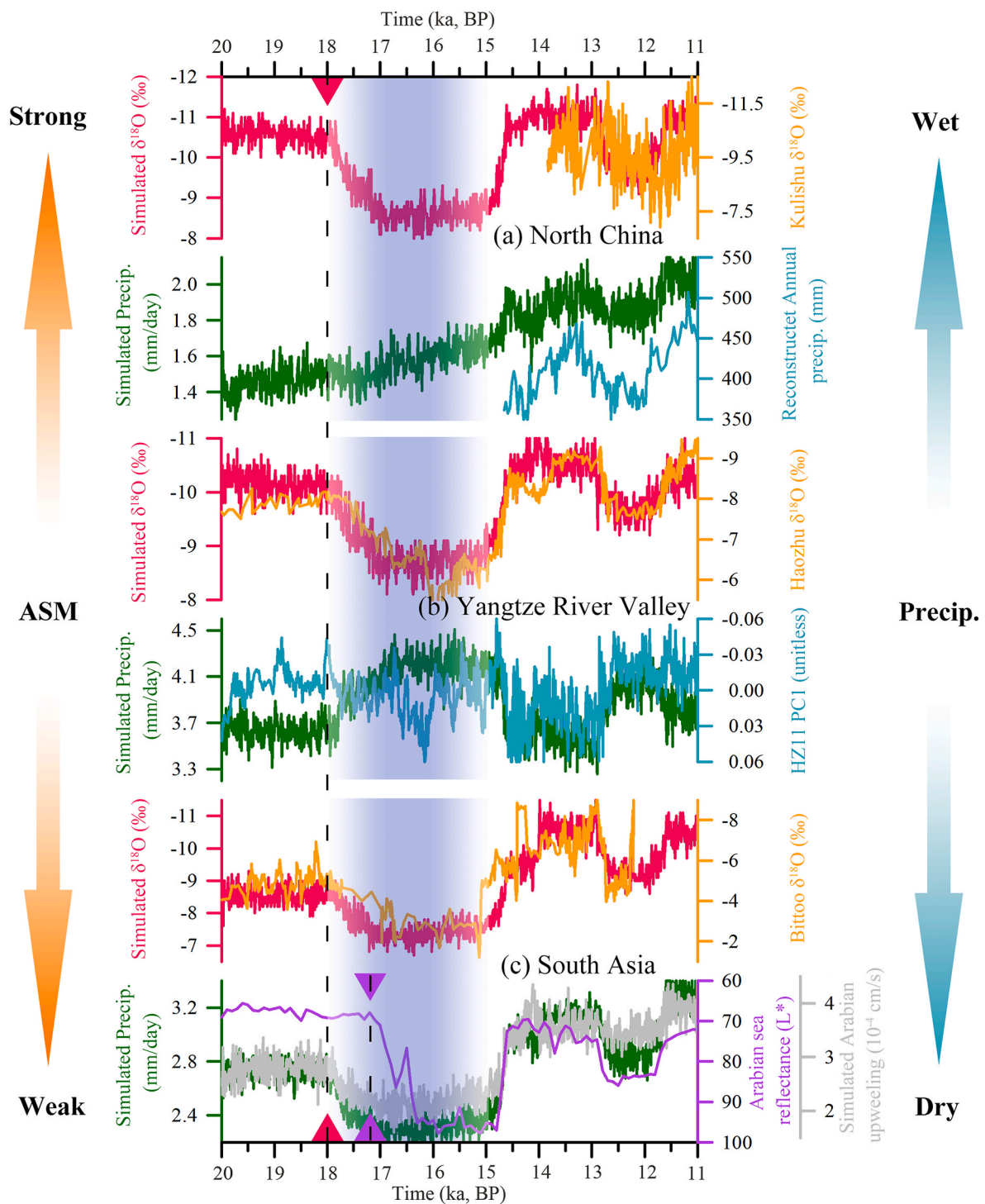


Fig. 2. Comparison the Asian summer monsoon spatial pattern between the reconstructed paleo-monsoon and paleo-rainfall records with model results (iTRACE, He et al., 2021) during the H1 event. Dark blue bars indicate the H1 event. (a) North China (36°N–43°N, 105°E–120°E): simulated speleothem $\delta^{18}\text{O}$ (pink) and annual precipitation (green) in iTRACE, reconstructed $\delta^{18}\text{O}$ in Kulishu cave (orange, Ma et al., 2012) and annual precipitation in Gonghai lake (dark cyan, Chen et al., 2015). (b) Yangtze River Valley (28°N–35°N, 105°E–120°E): simulated speleothem $\delta^{18}\text{O}$ (pink) and annual precipitation (green) in iTRACE, reconstructed $\delta^{18}\text{O}$ in Haozhu cave (orange, Zhang et al., 2018) and summer monsoon precipitation index inferred from the PC1 of Mg/Ca, Sr/Ca and Ba/Ca records in Haozhu cave (dark cyan, Zhang et al., 2018). (c) South Asia (10°N–25°N, 70°E–95°E): simulated speleothem $\delta^{18}\text{O}$ (pink) and annual precipitation (green) in iTRACE, reconstructed $\delta^{18}\text{O}$ in Bittoo cave (orange, Kathayat et al., 2016), simulated Arabian Sea upwelling (gray, 23°N–24.3°N, 62.8°E–65°E), and reconstructed Arabian Sea sediment total reflectance as a proxy for SASM wind strength (pink, Deplazes et al., 2013). Red triangle and dash line indicate the onset of H1 in simulated EASM and paleo-monsoon records around 18ka and about 700y delay of SASM onset in Arabian Sea sediment total reflectance record shown by pink triangle and dash line.

Table 1

Summary of the orbital parameters (Berger and Loutre, 1991) for the Heinrich Event-Orbital experiment simulations.

Table 1 The top and base ages of six H events (Sanchez Goñi and Harrison, 2010) and the orbital parameters (Berger and Loutre, 1991) used in the simulations in this study.

	Age (ka)		Eccentricity	Obliquity	Longitude of perihelion
PI	0		0.01724	23.45	101.37
H1	Top	15.6	0.02000	23.62	179.38
	Base	18			
H2	Top	24.3	0.01827	22.41	40.49
	Base	26.5			
H3	Top	31.3	0.01627	22.39	296.86
	Base	32.7			
H4	Top	38.3	0.01450	23.43	193.72
	Base	40.2			
H5	Top	47	0.01462	24.42	66.09
	Base	50			
H6	Top	60.1	0.01934	23.04	242.17
	Base	63.2			

and Lan, 2012):

$$P - E = -\langle \omega \partial_p q \rangle - \langle \vec{V}_H \cdot \nabla q \rangle + Res \quad (1)$$

Where P and E are the precipitation and evaporation rate at the surface, ω is the vertical velocity at a constant pressure coordinate, q is the specific humidity, $\vec{V}_H$ is the horizontal wind, and Res is a residual term including transient eddy effects and a contribution due to changes in surface moisture transport by the surface pressure gradient (Seager et al., 2010; Chou and Lan, 2012; D'Agostino and Lionello, 2020). The contribution of the residual term to precipitation is typically ignored due to its generally minor impact. The angle brackets indicate vertical integral over the atmospheric column from the surface to 100 hPa. The local net precipitation on the left side of Eq. (1) contributes to the remote moisture process (on the right side) that includes vertical moisture advection, horizontal moisture advection and residual terms. The precipitation difference between the H events and PI can be derived from Eq. (1) as follows:

$$\delta P - \delta E = -\delta \langle \omega \partial_p q \rangle - \delta \langle \vec{V}_H \cdot \nabla q \rangle + \delta Res \quad (2)$$

To further distinguish the relative contributions of changes in specific humidity and changes in atmospheric circulation to changes in water vapor transport and hence precipitation, changes in both horizontal and vertical moisture transports are decomposed into: 1) thermodynamic contribution (TH, Eq. (3)) caused by changes in specific humidity with unchanged circulation; 2) dynamic contribution (DY, Eq. (4)) caused by atmospheric circulation changes with unchanged specific humidity, and non-linear (NL, Eq. (5)) effects from simultaneous changes in atmospheric circulation and specific humidity (Sun et al., 2021, 2023).

$$TH = -\langle \vec{V}_H \cdot \nabla \delta q \rangle - \langle \omega \partial_p \delta q \rangle = -\frac{1}{g\rho_w} \int_{ps}^{pt} \vec{V}_H \cdot \nabla \delta q dp - \frac{1}{g\rho_w} \int_{ps}^{pt} \omega \partial_p \delta q dp \quad (3)$$

$$DY = -\langle \delta \vec{V}_H \cdot \nabla q \rangle - \langle \delta \omega \partial_p q \rangle = -\frac{1}{g\rho_w} \int_{ps}^{pt} \delta \vec{V}_H \cdot \nabla q dp - \frac{1}{g\rho_w} \int_{ps}^{pt} \delta \omega \partial_p q dp \quad (4)$$

$$NL = -\langle \delta \vec{V}_H \cdot \nabla \delta q \rangle - \langle \delta \omega \partial_p \delta q \rangle = -\frac{1}{g\rho_w} \int_{ps}^{pt} \delta \vec{V}_H \cdot \nabla \delta q dp - \frac{1}{g\rho_w} \int_{ps}^{pt} \delta \omega \partial_p \delta q dp \quad (5)$$

where ρ_w is the water density, g is gravitational acceleration, ps is the surface pressure and pt is equal to 100 hPa. Furthermore, the DY also can be further decomposed into three terms as:

$$DY = -\langle \delta u \frac{\partial u}{\partial x} \rangle - \langle \delta v \frac{\partial v}{\partial y} \rangle - \langle \delta w \partial_p q \rangle \quad (6)$$

Where δu , δv , and δw the zonal wind, meridional wind, and vertical pressure-velocity anomalies, respectively. The three terms on the right side in Eq. (6) denote dynamic contribution to the zonal, meridional and vertical moisture transports changes, abbreviated as XD, YD and WD, respectively.

2.3. Definition of Asian monsoon index

Given the pronounced seasonal variation of the monsoons, it makes sense to assess the strength of the monsoons using the wind field's seasonal variation amplitude. Thus, the SASM strength index (SASMI) and EASM strength index (EASMI) are defined as an area-averaged dynamical normalized seasonality wind at the 850 hPa wind field in JJA within the SASM domain (5°-22.5°N, 35°-97.5°E) and EASM domain (10°-40°N, 110°-140°E), respectively (Li and Zeng, 2002). The dynamical normalized seasonality monsoon index is given by the following:

$$\sigma_{m,n} = \frac{\|\vec{V}_1 - \vec{V}_{m,n}\|}{\|\vec{V}\|} \quad (7)$$

Where the $\vec{V}_1$ and $\vec{V}$ are the January climatological wind vector and the mean of January and July climatological wind vectors, respectively, and $\vec{V}_{m,n}$ is the monthly wind vector for the m th month of the n th year. The norm $\|\vec{V}\|$ is defined as:

$$\|\vec{V}\| = \left(\iint_S |\vec{V}|^2 dS \right)^{\frac{1}{2}} \quad (8)$$

where S is the domain of integration. All the monsoon indexes are smoothed with 11 years in this study.

2.4. Causal analysis

As a correlation between two series does not imply causation (Liang, 2014), to determine the causal relationship between monsoon precipitation, ASM intensity, and migration of ITCZ, the Liang-Kleeman information flow approach is utilized based on the information theory and first principles (Liang and Kleeman, 2005; Liang, 2014, 2016). For two given time series x_1 and x_2 , the maximum likelihood estimate of information flow from x_2 to x_1 per unit time can be expressed as follows:

$$T_{2 \rightarrow 1} = \frac{C_{11}C_{12}C_{2,d1} - C_{12}^2C_{1,d1}}{C_{11}^2C_{22} - C_{11}C_{12}^2} \quad (9)$$

$$C_{ij} = \overline{(x_i - \bar{x}_i)(x_j - \bar{x}_j)} \quad (10)$$

$$\dot{x}_{j,n} = (x_{j,n+k} - x_{j,n}) / k\Delta t \quad (11)$$

Where C_{ij} is the sample covariance between x_i and x_j , $C_{i,dj}$ is the sample covariance between x_i and the Euler forward finite difference of x_j , Δt is the time step and $k \geq 1$ (integer). The flow in the opposite direction $T_{1 \rightarrow 2}$ can be directly written by exchanging the subscripts 1 and 2. According to this theory, if $T_{1 \rightarrow 2} = 0$, X_2 does not cause X_1 ; if it is not equal to zero, it is causal (statistical significance test also performed). The strength of the causality from X_2 to X_1 is determined by $|T_{2 \rightarrow 1}|$ (Liang, 2016). This method has been recognized and applied to the cause-effect analysis of El Niño and Indian Ocean Dipole (Liang, 2014), CO₂ and global warming (Stips et al., 2016) and Arctic Sea ice and Tibetan plateau snow (Zhang et al., 2023).

3. Results and discussion

3.1. Different effect of orbital forcing on ASM during six heinrich events

The difference of the seasonal and latitudinal insolation distribution between the six H events and the PI is shown in Fig. 3. During H1, H3, H4, and H6, the Earth receives more insolation during boreal summer and less during boreal winter. This is a typical feature caused by change in precession. Increased summer insolation and reduced winter insolation in the Northern Hemisphere (NH) lead to enhanced seasonal insolation contrast in the NH, and it is the opposite in the Southern Hemisphere (SH) (Fig. 3a, c, d, f). During H5 (Table 1, Fig. 3e), the increased obliquity induces an asymmetrical insolation anomaly with respect to the Equator. This leads to enhanced insolation at high latitudes and reduced insolation at middle and lower latitudes during local summer, with the pattern reversed during local winter. Consequently, the inter-hemispheric (NH-SH) insolation contrast is intensified during boreal summer and diminished during boreal winter. The situation is the opposite during H2 due to a much smaller obliquity. The changes of precession and obliquity lead to a positive insolation gradient between 30°N and 30°S in boreal summer during the H1, H3-6 events, but a negative one during the H2 event (Fig. 3b).

In response to insolation changes, the increase of surface air temperature (SAT) in boreal summer is larger in the NH than in the SH during the H1, H3-6 events (Fig. 4a, c, d, e, f), which is in agreement with a positive insolation gradient between 30°N and 30°S. This in turn results in a weakening of the Asiatic Low and the Mascarene High with the former experiencing a more significant decrease than the latter, leading to an amplified meridional land-sea pressure gradient (Fig. 4a, c, d, e, f). Consequently, this enhances the northward cross-equatorial moisture transport along the Somali and North Australian regions towards the Asian continent interior, reinforcing the strength of the SASM and EASM. The intensification of ASM, along with the northward shift of the ITCZ (Fig. 5a, c, d, e, f), defined by changes in the zonal-mean maximum precipitation centroid between 20°S and 20°N (Donohoe et al., 2013), results in increased precipitation over the EASM region and the southern edge of the Tibetan Plateau, while reducing precipitation over the SASM region (Fig. 5a, c, d, e, f). This leads to significant spatial variations in precipitation across the ASM region. In addition, the

pronounced warming over the Asian continent, accompanied by minimal warming and even cooling over the North Pacific, means an increased east-west thermal contrast between land and sea. This coincides with a strengthening of the subtropical high in the western North Pacific and a weakening of the Asiatic Low (Fig. 4a, c, d, e, f). As a result, moisture transport from the southeast along the western edge of the Western North Pacific Subtropical High towards North China increases, leading to increased precipitation in North China (34°N–43°N, 105°E–120°E) (Fig. 5a, c, d, e, f). The H2 event, on the other hand, shows an opposite scenario. The cooling in the NH and the warming in the SH coincide with strengthening of the Asiatic Low and a weakening of the Mascarene High and the Western North Pacific Subtropical High, which reduces the meridional and zonal land-sea thermal contrast (Fig. 4b). This leads to a decrease in the strength of the SASM and the EASM, resulting in an anomaly southward equatorial moisture transport along the Somali and northern Australian region to the ocean (Fig. 5b). The weakening of ASM intensity combined with southward shift of the ITCZ during the H2 event result in widespread drought in North China, South China (22°N–28°N, 105°E–120°E) and SASM region (Fig. 5b). It is noteworthy that contrary to North China, South China and the SASM region, precipitation over the Yangtze River Valley (28°N–34°N, 105°E–120°E) increases during H2 (Fig. 5b), leading to significant spatial variations of precipitation in the ASM region. In summary, our model results show that changes in insolation alone can induce spatial discrepancies in precipitation over the ASM region during the H events.

3.2. Decomposing the physical processes controlling regional precipitation

To gain a comprehensive understanding of physical mechanisms underneath governing these significant spatial variations in precipitation within the ASM region, two cases with markedly different insolation pattern are selected to conducted area-averaged moisture budget analysis over the SASM and EASM domains, as indicated by the enclosed dash red box in Figs. 6 and 7. They are the H2 event with a negative 30°N–30°S insolation gradient during boreal summer and the H6 event with a positive one. As H6 shares a similar insolation gradient (Fig. 3) and a broadly similar climate response (Figs. 4 and 5) with H1, H3, H4 and H5 in terms of the hemisphere insolation contrast, these H events also share a similar result of the moisture budget analysis (not shown).

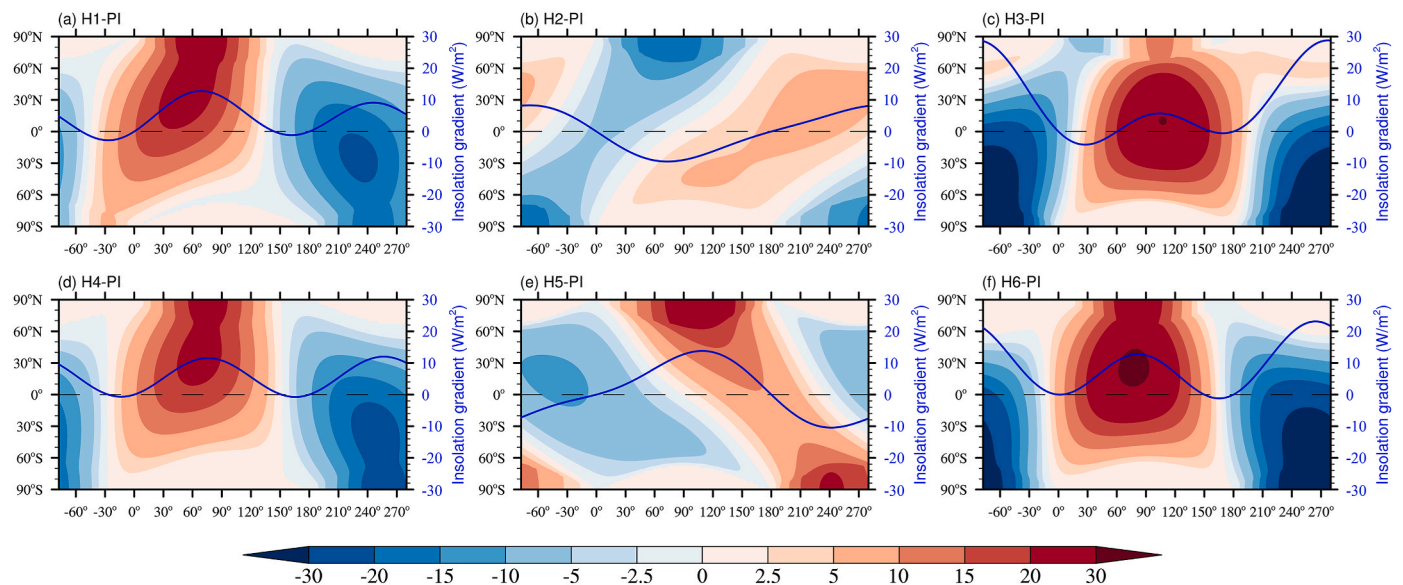


Fig. 3. Latitudinal-seasonal distribution of insolation differences (W/m^2) between six H events and PI. X-axis indicates the true longitude of the Sun over a year, with 0°, 90°, 180° and 270° corresponding to spring equinox, summer solstice, fall equinox and winter solstice, respectively. Insolation is calculated from the long-term variations of eccentricity, precession and obliquity (Berger and Loutre, 1991). The blue line indicates the boreal summer insolation gradient between 30°N and 30°S; the black dash line indicates the equator.

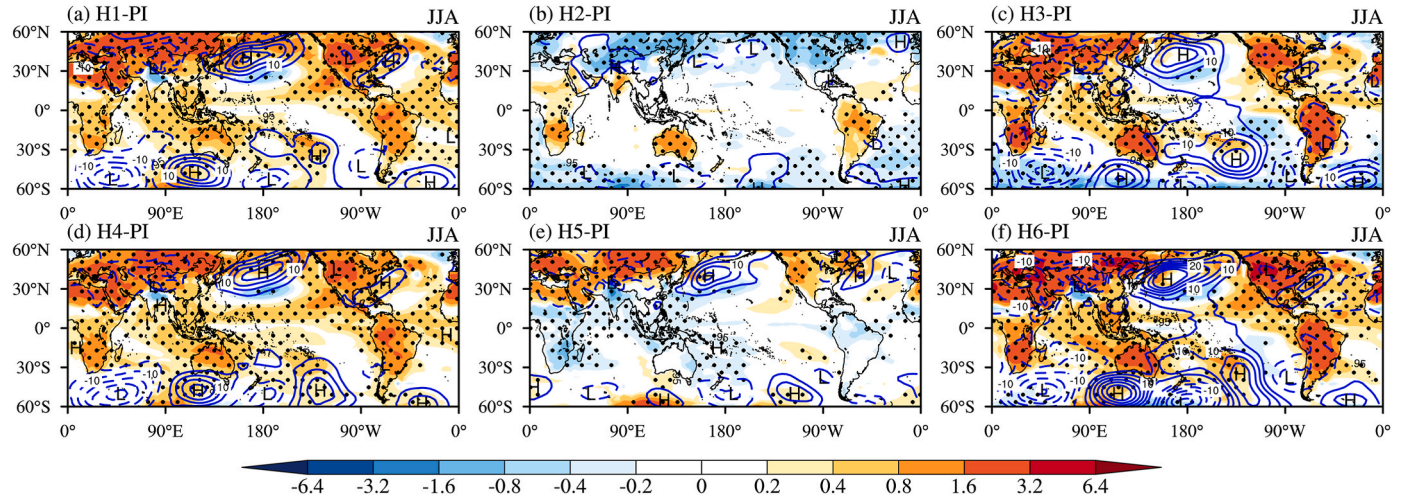


Fig. 4. Differences of boreal summer surface air temperature (shading, stippled for 90% significance level based on Student's t-test, unit: °C) and sea level pressure (contour, solid line for the positive and dash line for the negative, interval: 1 hPa) between six H event (a–f) and PI period. High- and Low-pressure centers are indicated with H and L, respectively.

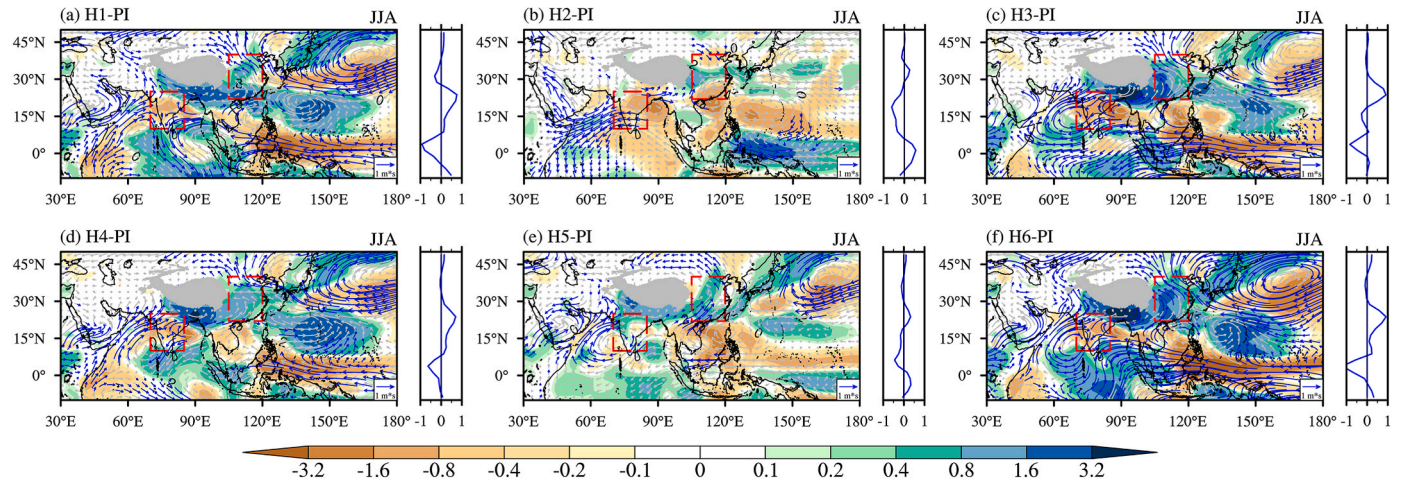


Fig. 5. Same as Fig. 5, but for net precipitation (precipitation minus evaporation, shading, unit: mm/day) and 850 hPa wind (vector, unit: m/s). The right panel shows changes in zonal-mean net precipitation (unit: mm/day) between 40°E–140°E. The blue vectors indicate the anomalies that are significant at the 90% level based on Student's t-test. The red boxes outline the SASM (10°N–25°N, 70°E–85°E) and EASM (22°N–43°N, 105°E–120°E) regions, the gray shaded area outlines the Tibetan Plateau.

Precipitation change is balanced by changes of the evaporation, dynamic, thermodynamic, nonlinear components (Eq.(2)–(5)) and residual term. The dynamic component can be further decomposed into the zonal, meridional, and vertical components of moisture transport (Eq. (6)). As compared to PI, the reduced SASM during H6 is primarily controlled by dynamic processes, with contrasting contributions from thermodynamic process and evaporation that result in enhanced H6 SASM precipitation (Fig. 6a). As precipitation decreases and evaporation increases within the SASM region, the net SASM precipitation (P-E) consequently decreases during H6 (Fig. 6b). This reduction in net precipitation leads to regional drought in the SASM region, primarily stemming from a decrease in moisture transport controlled by dynamic processes (Fig. 6a, b, c). Conversely, an increase in moisture transport controlled by thermodynamic processes leads to a wet SASM condition during H6 (Fig. 6d). Furthermore, decomposing the dynamic component into contributions from the zonal, meridional, and vertical components of moisture transport shows that the primary contribution arises from vertical moisture transport (Fig. 6g), followed by zonal component (Fig. 6e), with the meridional component playing a minor role (Fig. 6f). In contrast, the EASM region experiences a wetter condition during H6

primarily due to a considerable rise in precipitation, surpassing a modest increase in evaporation (Fig. 6a and b). This wetness is predominantly attributed to dynamic increase in precipitation (Fig. 6c), whereas the thermodynamic component exhibits negligible contribution to precipitation changes (Fig. 6a–d). Furthermore, consistent with the SASM region, dynamic enhancement of EASM precipitation during H6 is predominantly governed by the increased contribution due to vertical component (Fig. 6g), with the zonal component reducing the precipitation (Fig. 6e) and the meridional component enhancing it in North China (Fig. 6f).

During the H2 event, both the SASM and EASM regions typically experience a significant drought attributed to a strong reduction in precipitation (Fig. 7a and b), aligning with a reduced boreal summer insolation. The drought in both regions is largely due to decrease in the dynamic contribution to precipitation (Fig. 7a–c). Decomposition of the dynamic component highlights the prevalent influence of the vertical dynamic component in both the SASM and EASM regions (Fig. 7a–g). Nevertheless, it is noteworthy that the meridional component's contribution to precipitation decrease in North China should not be overlooked (Fig. 7f). This result is in line with what has been found during

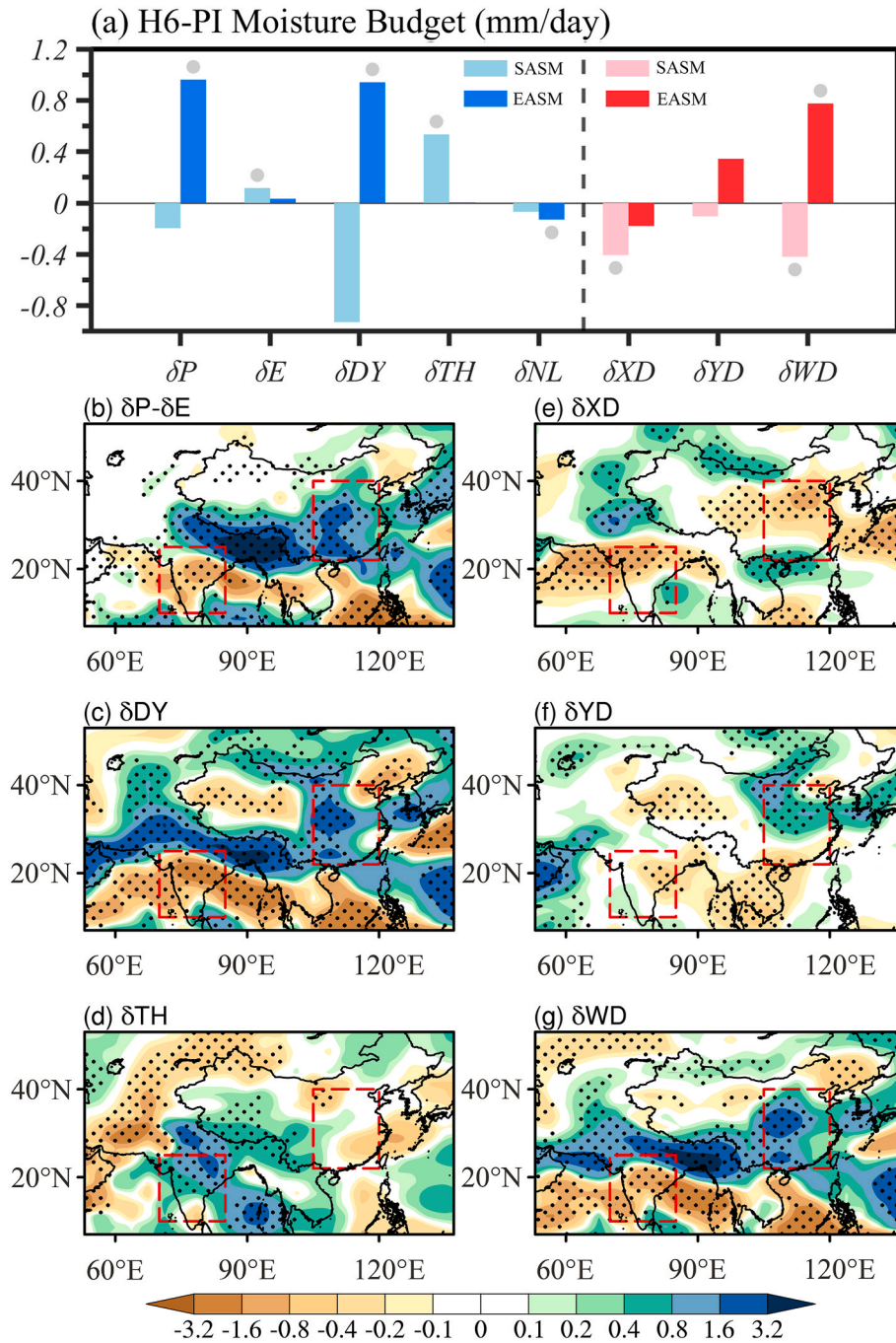


Fig. 6. Moisture budget analysis of summer precipitation changes (mm/day) over the Asian summer monsoon region between H2 and PI. (a) Decomposition of precipitation changes in the SASM and EASM regions. δP : precipitation changes; δE : evaporation changes; δDY : dynamic component; δTH : thermodynamic component; δNL : nonlinear term; δXD , δYD and δWD : zonal, meridional and vertical contributions in the dynamic component, respectively. Gray dots above the bars indicate the terms that are statistically significant at 90% confidence level based on Student's t-test. (b, c, d) Spatial patterns of net precipitation changes ($\delta P - \delta E$), dynamic component (δDY) and thermodynamic component (δTH), respectively. (e, f, g) zonal, meridional and vertical dynamic component of moisture advection, respectively. Areas with a significance above the 90% level are stippled. The red boxes outline the SASM (10°N–25°N, 70°E–85°E) and EASM (22°N–43°N, 105°E–120°E) regions used in (a).

the H6 period (YD component in Figs. 6a and 7a). Despite the widespread drought in the ASM region as expected from the weakened SASM and EASM intensity and the southward shift of ITCZ related to boreal summer insolation decrease, the YRV encounters abnormal wetness during H2, emphasizing the spatial heterogeneity of precipitation changes (Fig. 7b). This abnormal wetness is primarily attributed to increased precipitation due to the dynamic component (Fig. 7c), particularly the vertical component (Fig. 7g) and the zonal component (Fig. 7e).

The moisture budget analysis for the H2 and H6 events shows that the dynamic component especially the vertical dynamic component dominates the precipitation changes over the SASM and EASM regions, while the thermodynamic component plays a minor role. The SASM region is also notably influenced by the zonal dynamic component, while the EASM region, particularly the North China, is also influenced by the meridional component.

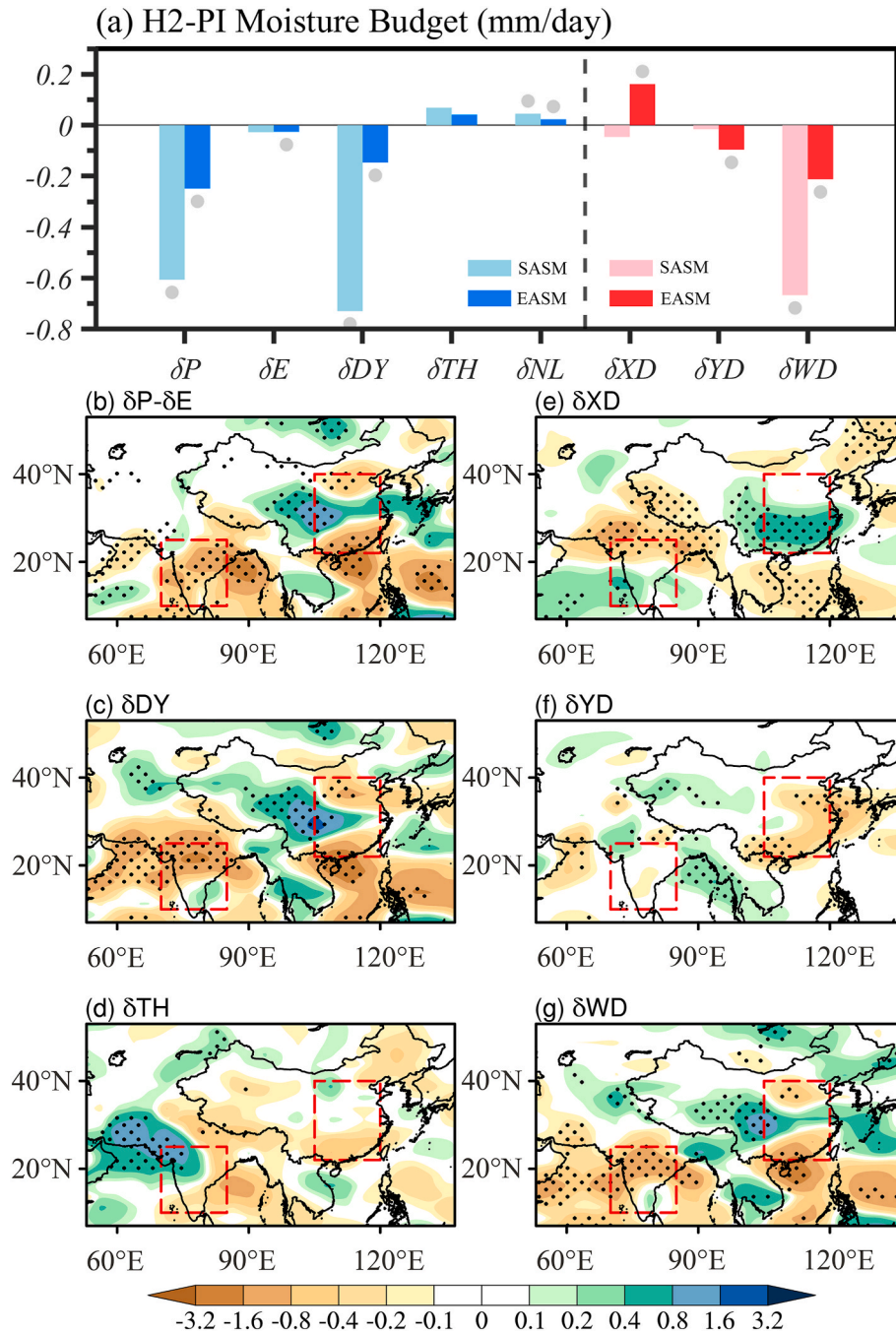


Fig. 7. Same as Fig. 6, but for the difference between H6 and PI.

3.3. Precipitation changes related with large-scale water vapor transport and its convergence

To explain the insolation-induced precipitation changes during the H2 and H6 events as shown in Section 3.2, the associated large-scale atmospheric circulations and water vapor transport are analyzed. Fig. A2 displays the climatological mean distribution of water vapor transport fluxes and convergence/divergence pattern during boreal summer under the PI condition. It shows an organized divergence zone along the Somalia's coast, together with abundant cross-equatorial water vapor carried by the ASM flow from southern Indian Ocean and Arabian Sea to the Indian subcontinent, where it convergences to increase precipitation. Subsequently, the water vapor continuously transports moisture from west to east, crossing the Bay of Bengal,

Indochina Peninsula, South China sea, and then shifting northward towards the EASM region (Huang et al., 2017). In addition to the moisture originating from the SASM region, the transport of moisture from the northwest Pacific Ocean along with flank of Western North Pacific Subtropical High and cross-equator flow straddling 105°E – 150°E, also play an important role in the EASM region (Huang et al., 2017). The three branches of moisture transport converge over the EASM region to produce precipitation there. Therefore, as generally acknowledged (Wang et al., 2001a,b; Ding and Chan, 2005; Wu, 2017; Huang et al., 2017), the SASM exhibits more tropical characteristics, whereas the EASM displays characteristics of both tropical and subtropical influences. Consequently, the EASM should not simply be regarded as eastward and northward extension of the SASM.

During the H6 event, the amplification of the meridional land-sea

pressure contrast in response to a positive insolation gradient during boreal summer intensifies the northward cross-equatorial moisture transport from the Somali region to the northern Indian subcontinent, accompanied by an anomalous anticyclone over the Arabian Sea (Fig. 8a). However, an anomalous cyclone over the Bay of Bengal, contrary to the climatological westerly moisture transport, leads to a

reduction of moisture transport across the southern Indian region (Fig. 8a). In general, the increase of moisture transport across the whole SASM region (red box) doesn't result in increased precipitation, primarily due to the anomalous moisture divergence over the whole SASM region (Fig. 8a). This anomalous moisture divergence corresponds well with the northward movement of ITCZ, and anomalous downdraft

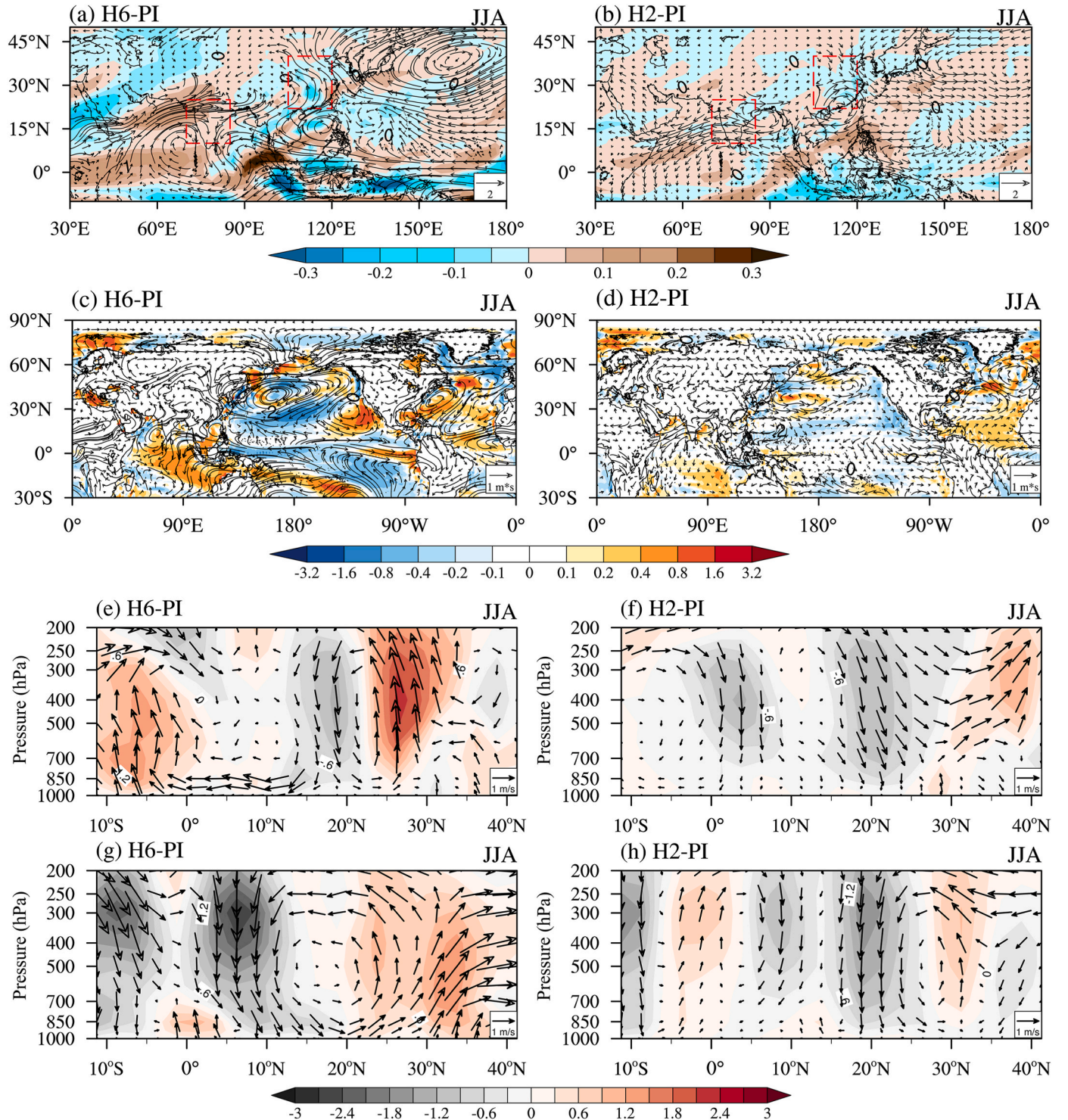


Fig. 8. Boreal summer climatology differences for the H6-PI (left panel) and H2-PI (right panel). (a, b) 850 hPa moisture flux divergence (shading, unit: $10^{-7} \text{ g/cm}^2\text{s}^*\text{hPa}$) and moisture flux (vectors, unit: $\text{g/cm}^2\text{s}^*\text{hPa}$). (c, d) Sea surface temperature zonal amplitude (shading, unit: $^{\circ}\text{C}$) and 850 hPa wind (vectors, unit: m/s). (e, f) Meridional-vertical distribution of regional Hadley cell circulation (vector, v-wind vs. $-100^{\circ}\omega$) and $-100^{\circ}\omega$ wind (shading, unit: Pa/s) average between 70°E – 85°E over the SASM region (left red box in (a)). (g, h) Same as (e, f) but for the average between 105°E – 120°E over the EASM region (right red box in (a)).

occupy the whole SASM region (Fig. 8e). These anomalous anticyclone over the Arabian Sea and cyclone over the Bay of Bengal are potentially triggered by a positive Indian Ocean Dipole (Fig. 8c), suggesting the influence of Indian Ocean SST on the SASM strength and precipitation. Over the EASM region, the moisture transport from the Bay of Bengal, the South China sea, and the northwestern Pacific Ocean is enhanced during H6 as compared to PI. This is due to an enhancement of the eastward moisture transport by the SASM, of the northward moisture transport along the western side of the anomalous anticyclone over South China Sea, and of the westward moisture transport along the western flank of the abnormally stronger Western North Pacific Subtropical High (Fig. 8a). The increased moisture transport, coupled with anomalous moisture convergence (Fig. 8a), contributes to increased precipitation over the EASM region during H6. The anomalous moisture convergence is accompanied by the northward shift of the ITCZ and the presence of anomalous updraft (Fig. 8e), spanning across the entire EASM region.

The opposite case occurs during the H2 event. The negative insolation gradient during boreal summer weakens the meridional land-sea pressure contrast, resulting an anomaly southward cross-equatorial moisture transport from the NH to the SH. The moisture transport and moisture divergence over the whole SASM region decrease, leading to less precipitation, which is accompanied by southward shift of ITCZ and anomalous downdraft over the whole SASM region (Fig. 8b, f). Over the EASM region, the enhanced southerly warm, moist vapor transport at the western edge of the subtropical Western North Pacific and the northerly cold, dry vapor transport along the western edge of the mid-latitude cyclone anomaly appear to meet and converge around the Yangtze River Valley, and cause more precipitation there (Fig. 8b). This convergence is accompanied by anomalous updraft around 30°N (Fig. 8h). However, the anomalous downdraft on both sides of the Yangtze River Valley, accompanied by moisture divergence, results in less precipitation over North and South China (Fig. 8h). It's noteworthy that an anticyclone anomaly is present over the subtropical Western North Pacific, accompanied by a cyclone anomaly over Japan and Korea (Fig. 8b). This circulation pattern captures a meridional wave train from the subtropical Western North Pacific to the north of Japan, known as the Pacific-Japan (PJ) pattern as previously proposed by Nitta (1987) and Huang and Wu (1989) and this pattern (Fig. 8d) highlights the important role of western tropical Pacific in the linkage of EASM and ENSO.

The analysis of the large-scale moisture transport and its convergence shows that, regardless of whether water vapor transport strengthens or weakens, precipitation in the SASM region always decreases during the H2 and H6 periods, which is primarily attributed to water vapor divergence and the presence of anomalous downdrafts over the SASM region. The increased precipitation during H2 across the Yangtze River Valley results from an anticyclone anomaly over the subtropical Western North Pacific and a cyclone anomaly over Japan and Korea (Fig. 8d). This P-J pattern results in the convergence of anomalous northerly cold and dry air with southerly warm and moist air around the Yangtze River Valley.

3.4. The relationship between the ASM strength, precipitation and ITCZ

To investigate the relationship between the ASM strength and summer precipitation in response to orbital forcing, the correlation coefficient between the ASM intensity index and local precipitation is first calculated (Fig. 9). There is significant positive correlation between the SASMI and EASMI (Fig. 9a), indicating consistent changes in the large-scale spatial circulation (low-level wind field) within the ASM region. These changes are attributed to in-phase variations in the meridional (Eurasia Continent-Southern Indian Ocean) and zonal (Eurasia continent- Western Pacific Ocean) land-sea thermal contrast and the migration of the Intertropical Convergence Zone driven by orbital forcing. However, there is huge spatial heterogeneity in the relationship between the ASM intensity and local precipitation. Over the SASM region, there seems to be no significant relationship between the SASMI and SAMP on orbital time scale (Fig. 9b), in contrast to modern observations that show a significant positive relationship on interdecadal time scale (Li and Zeng, 2005). This positive relationship leads to the widespread use of SAMP as an indicator of SASMI in both modern and paleoclimate studies (Gadgil, 2003; Kathayat et al., 2016). However, it's important to acknowledge that this relationship weakened after 1979, attributed to global warming, aerosol effects, deforestation, and a negative-to-positive phase transition of the interdecadal Pacific oscillation (Huang et al., 2020). Consequently, caution should be taken when using SAMP to indicate SASMI, particularly under different climate backgrounds and time scales. Over the EASM region, a weak positive relationship generally exists between the EASMI and EASP (Fig. 9c), despite substantial discrepancies in precipitation across different locations. The different relationship between the ASM intensity and local

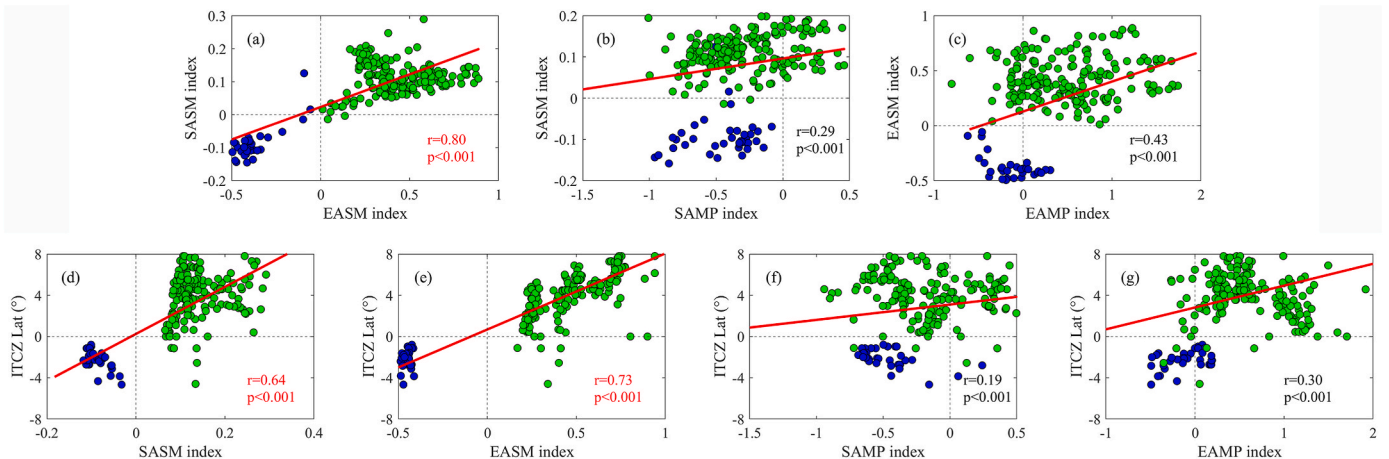


Fig. 9. Scatterplots and linear regression for the 11-point moving mean anomalies of boreal summer monsoon index, net precipitation (unit: mm/day), and ITCZ location (latitude). (a) SASM index and EASM index, (b) SASM index and SASM net precipitation, (c) EASM index and EASM net precipitation, (d) ITCZ location and SASM index, (e) ITCZ location and EASM index, (f) ITCZ location and SASM net precipitation and (g) ITCZ location and EASM net precipitation. The analysis is based on the last 60-year model outputs. Blue scatter is for the H2-PI anomalies and green scatter for the H1, 3, 4, 5, 6-PI anomalies. The r in each panel is the correlation coefficient with the significance level p . The ITCZ location anomaly is defined by changes in zonal-mean maximum precipitation centroid between 20°S and 20°N (Donohoe et al., 2013). The precipitation is interpolated to 1° grid over the tropical to allow the precipitation centroid to vary at increments smaller than the grid spacing.

Table 2

Information flow between ITCZ, SASM, EASM and precipitation variation

Table 2 Information flow between ITCZ, SASM, EASM and precipitation variation. The units time step is 1/year. Causation significant at the 90% confidence level with an absolute IF larger the 0.1 nat/year is shown in bold. The '± errors' represents the ±90% confidence.

Information flow direction (X2→X1)	Information flow (nat/year) (X2→X1)	Relative information flow (%) (X2→X1)	Information flow direction (X1→X2)	Information flow (nat/year) (X1→X2)	Relative information flow (%) (X1→X2)
SASM→SAMP	−0.270 ± 0.236	3	SAMP→SASM	−0.182 ± 0.106	1
SASM→EAMP	−0.121 ± 0.202	13	EAMP→SASM	0.103 ± 0.124	5
EASM→SAMP	−0.098 ± 0.173	4	SAMP→EASM	−0.041 ± 0.052	4
EASM→EAMP	−0.051 ± 0.220	2	EAMP→EASM	0.056 ± 0.089	−13
SASM→EASM	0.393 ± 0.223	44	EASM→SASM	0.369 ± 0.249	38
ITCZ→SAMP	0.002 ± 0.102	1	SAMP→ITCZ	0.103 ± 0.050	−5
ITCZ→EAMP	−0.018 ± 0.167	1	EAMP→ITCZ	0.197 ± 0.110	−5
ITCZ→SASM	0.396 ± 0.217	45	SASM→ITCZ	0.282 ± 0.113	29
ITCZ→EASM	0.366 ± 0.264	36	EASM→ITCZ	0.286 ± 0.138	22

The units time step is 1/year. Causation significant at the 90% confidence level with an absolute IF larger the 0.1 nat/year is shown in bold.

The '± errors' represents the ±90% confidence level.

precipitation changes can be partly attributed to the location of the ITCZ (Fig. 9d, e, f, g). The SASM is part of the tropical monsoon system and essentially lies between the seasonal migration boundaries of the ITCZ (An et al., 2015). During H2, the maximum ITCZ location anomaly situates over the Southern Hemisphere, while during H6, it resides on the north side of the SASM region, both anomaly situations contributing to a precipitation deficit over the SASM region (Fig. 5b, f & Fig. 8e and f). However, the EASM is influenced not only by the migration of ITCZ in South China but also by southeastward moisture transport along the western edge of the Western North Pacific subtropical high to the Yangtze River Valley and North China. The in-phase variations of the ITCZ location (or tropical monsoon) and subtropical Western North Pacific moisture transport result in a significant positive relationship between the East Asian Summer Monsoon Index (EASMI) and summer precipitation (EAMP) under orbital forcing (Fig. 9g).

Despite the observed significant relationship between the migration of the ITCZ, the ASM strength and local precipitation, it is important to note that causation cannot be conclusively established only based on the correlation analysis. The Liang–Kleeman information flow approach is further employed to determine the causal relationship between them as shown in Table 2. There exists significant mutual causal relationship between the migration of the ITCZ and the variation of the ASM. The higher information flow from the ITCZ to the SASM and EASM as compared to the reverse direction (SASM, EASM to ITCZ) suggests that the movement of the ITCZ forces the ASM variation, which is in line with the perspective that monsoons serve as a manifestation of the seasonal migration of the ITCZ (Gadgil, 2003). In turn, the onset of the Asian summer monsoon also subsequently forces the ITCZ northward movement. With the onset of the SASM, the intensified latent heat release from its rainfall may induce an upper-level wave pattern with anticyclone anomalies over the Northeast Asia and an anomalous cyclone over eastern China (Wu, 2017). This wave pattern propagates eastward along the westerly jet stream, establishing a connection between SASM and the EASM. In addition, this teleconnection between the SASM and EASM also can be reflected in zonal shift of the South Asian high, which dynamically affects the zonal extension of the Western North Pacific Subtropical High (Wei et al., 2019). Consequently, the significant mutual causal relationship between the strength of SASM and EASM, with the former exerting a greater information flow to the latter (Table 2).

4. Conclusions

The influence of orbital forcing on the spatial discrepancies in precipitation and the relationship between the EASM and SASM intensity during six H events are investigated by using simulations with the HadCM3 model. Moisture budget analysis is carried out to examine the physical processes governing the precipitation changes in response to

orbital forcing and the Liang–Kleeman information flow is employed to determine the causal relationship between the migration of the ITCZ, the ASM strength and local precipitation.

The results show that changes in insolation alone can induce spatial discrepancies in precipitation over the Asian monsoon region. During the H1, H3–6 events, the amplification of the land-sea pressure contrast in response to a positive insolation gradient during boreal summer intensifies moisture transport from the ocean to the Asian monsoon region. However, the ensuing moisture divergence, combined with anomalous downdrafts, results in decreased precipitation in the SASM region, and the converse scenario happens in the EASM region. During the H2 event, the increased precipitation across the Yangtze River Valley sharply contrasts the widespread drought over the ASM region. The precipitation increase across the Yangtze River Valley is attributed to the convergence of an anomalous northerly cold, dry vapor transport and an anomalous southerly warm, moist vapor transport, which are associated respectively with a cyclone anomaly over Japan and Korea and an anticyclone anomaly over the subtropical Western North Pacific. This P–J pattern could result from an anomalous cooling over the western tropical Pacific. This finding emphasizes the necessity of considering the impact of orbital forcing on spatial precipitation variations, despite the iceberg discharge (or freshwater forcing) being considered the main characteristic of H events.

Moisture budget analysis shows that precipitation changes over the SASM and EASM regions are dominated by the dynamic component, furthermore by the vertical component, while the thermodynamic component plays a minor role. In addition to the vertical dominance, the SASM region is also influenced by the zonal dynamic component, while the EASM region is more influenced by meridional dynamic component. It's worth noting that, although the significant positive relationship between the intensity of the EASM and SASM indicates in phase variation of SASM and EASM in response to the NHSI variations on millennial to orbital time scale, consistent with the Kutzbach's orbital monsoon hypothesis (1981), the EASM should not be simply regarded as an eastward and northward extension of the SASM. The former represents a tropical and subtropical monsoon system, while the latter is solely a tropical monsoon system. This notion warrants attention in paleomonsoon study and it could explain some observed paleoclimate inconsistency between the two monsoon regions. For example, a recently published speleothem $\delta^{18}\text{O}$ record from the Xianyun cave in South China indicates a gradual wetting trend from the early Holocene to late Holocene, in contrast to the prevalent drying trend in the SASM region (Zhang et al., 2021), emphasizing the divergent evolution of precipitation changes in the EASM and SASM regions during the Holocene. At last, our model results show a weak positive correlation between SASMI and SAMP in response to orbital forcing. This is contrary to modern observations, calling for caution when employing SAMP to reconstruct SASM intensity on orbital time scale and under different

climate backgrounds.

Our study provides an in-depth investigation of the dynamics of the response of the ASM circulation and precipitation to orbital forcing during six H events, and emphasizes the role of insolation in generating regional climate diversity during the H events. However, a thorough understanding of the spatial-temporal climate diversity during the H events and a comprehensive model-proxy comparison require considering also the effects of other forcing factors such as ice sheets, greenhouse gases, freshwater fluxes as well as vegetation feedback, investigations which will be carried out in future work.

CRediT authorship contribution statement

Ming-Qiang Liang: Conceptualization, Investigation, Methodology, Formal analysis, Visualization, Writing – original draft, Writing – review & editing. **Qiuzhen Yin:** Conceptualization, Funding acquisition, Methodology, Formal analysis, Supervision, Project administration, Writing – review & editing. **Yong Sun:** Methodology, Formal analysis, Writing – review & editing. **Chao Zhang:** Methodology, Formal analysis, Writing – review & editing. **André Berger:** Formal analysis, Writing – review & editing. **Anqi Lyu:** Formal analysis, Writing – review & editing. **Wei Liu:** Formal analysis, Writing – review & editing. **Zhipeng Wu:** Formal analysis, Writing – review & editing.

Declaration of competing interest

The authors state that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Appendix. Model performance

To assess the ability of HadCM3 to reproduce the present climate especially over the ASM region, we first compare the HadCM3 piControl simulation with the NOAA/CIRES/DOE-20CRv3 (Compo et al., 2011; Slivinski et al., 2019) reanalysis dataset for surface air temperature (SAT), sea level pressure (SLP), surface moisture static energy (MSE), precipitation and 850 hPa wind during boreal summer (Fig. A1). It can be seen that HadCM3 well reproduces the present-day ASM system. The stronger warming over the Asian continent than the adjacent oceans owing to differences in thermal inertia during boreal summer, is accompanied by a vast Asiatic Low centered at northern India and two huge subtropical high-pressure system over the South Indian Ocean (called Mascarene High) and North Pacific Ocean (called Western North Pacific Subtropical High) (Fig. A1d). The increased meridional thermal contrast leads to northward cross-equatorial atmospheric moisture transport from the southern Indian Ocean, and the increased zonal land-sea thermal contrast leads to southeastward atmospheric moisture transport from the subtropical Western North Pacific, both to the ASM region (Fig. A1f). The stronger surface warming over the ASM region, surface sensitive energy flux and advection of large-scale moisture transport lead to a band of maximum column-integrated moist static energy (MSE) in the ASM region (Fig. A1e), resulting in heavy summer monsoon precipitation (Fig. A1f, Wen et al., 2022). In general, these climatological spatial patterns and characteristics modelled by HadCM3 (Fig. A1d, e, f, right panel) broadly resemble the NOAA/CIRES/DOE-20CRv3 reanalysis dataset (Fig. A1a, b, c, left panel), although there are less precipitation in western Ghats and more precipitation in South China sea and Indonesian subcontinent region (Fig. A1f) in HadCM3 due to the “Double-ITCZ bias” which is one of the most outstanding errors in all previous generations of climate models (Tian and Dong, 2020). HadCM3’s performance over East Asia has also been shown to be among the best of many GCMs (Jiang et al., 2005) and it has been successfully used to simulate the Asian climate during the last interglacial (Lyu and Yin, 2022).

AppendixList of abbreviations

Abbreviation	Full name
H	Heinrich events
DO	Dansgaard-Oeschger events
IRD	Ice-rafted debris
AMOC	Atlantic Meridional Overturning Circulation
ITCZ	Intertropical Convergence Zone
ASM	Asian Summer Monsoon
SASM	South Asian Summer Monsoon
SASMI	South Asian Summer Monsoon strength index
SAMP	South Asian Summer Monsoon Precipitation
EASM	East Asian Summer Monsoon
EASMI	East Asian Summer Monsoon strength index
EAMP	East Asian Summer Monsoon Precipitation
TH	Thermodynamic contribution
DY	Dynamic contribution
NL	Non-linear contribution
NHSI	North Hemisphere Summer Insolation

Data availability

The model data to produce the main figures in this paper can be found at: <https://zenodo.org/records/13488210>.

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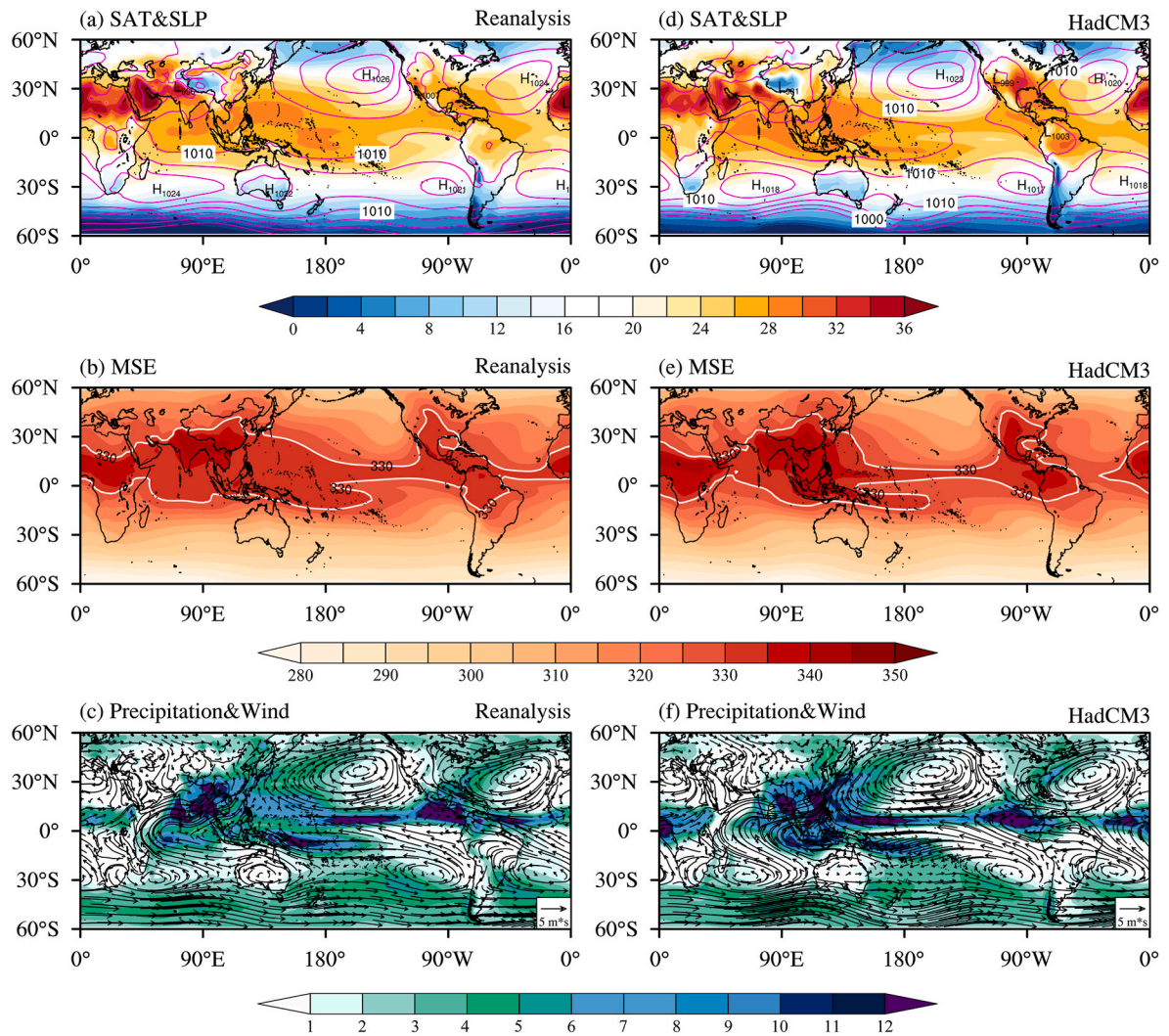


Fig. A1. Comparison of spatial distribution of boreal summer (JJA) hydroclimate over Asian summer monsoon region between reanalysis (Left panel, a-c) and HadCM3 piControl simulation (right panel, d-f). Reanalysis spans from 1836 to 1865 and is from NOAA/CIRES/DOE-20CrV3 (data source: https://www.psl.noaa.gov/data/gridded/data.20thC_ReanV3.html). (a, d) surface air temperature (SAT in °C, shading) and sea level pressure (SLP in hPa, contour, interval: 5 hPa); (b, e) surface moist static energy (MSE in K, shading) and the Northern Hemisphere monsoon regions are roughly outlined by the white contour line, where the MSE higher than 330 K (Wen et al., 2022). (c, f) net precipitation (shading, unit: mm/day) and 850 hPa wind (vector, unit: m/s).

Appendix. PI JJA moisture flux and divergence

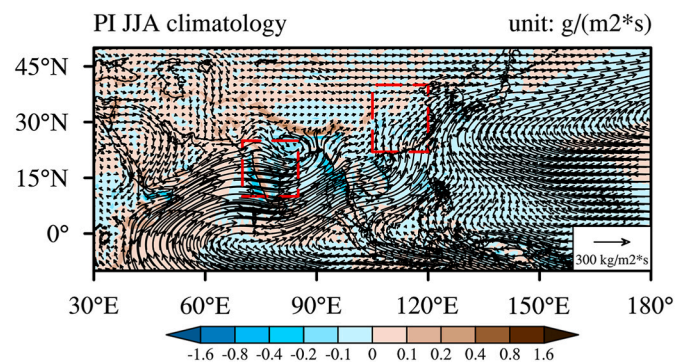


Fig. A2. Vertically integrated moisture flux convergence (shaded, unit: g/(m²*s)) and its transport (vector, unit: kg/(m²*s)) for the JJA season of the 20CrV reanalysis spans from 1836 to 1865 year. The red boxes outline the SASM (10°N–25°N, 70°E–85°E) and EASM (22°N–43°N, 105°E–120°E) regions.

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