DIE ERDE
Journal of the Geographical Society of Berlin

Vol. 150, No. 4 · Research article

Climate Variability in Indonesia from 615 ka to present: First Insights from Low-Resolution Coupled Model Simulations

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Manuscript submitted: 12 December 2018 / Accepted for publication: 16 August 2019 / Published online: 20 December 2019

Abstract

We analyse the dynamics of Indonesian waters using the results of a set of 13 time-slice experiments simulated by the CCSM3-DGVM model. The experiments were carried out to study global climate variability between and within the Quaternary interglacials of Marine Isotope Stages (MIS) 1, 5, 11, 13, and 15. During boreal summer (June-July-August-September), in most of Indonesia, seasonal surface temperature anomalies can largely be explained by local insolation anomalies induced by the astronomical forcing. However, for some time slices, climate feedbacks may modify the surface temperature response in Indonesia, most pronounced in open water close to the Indian and Pacific Oceans. The warmest boreal summer sea-surface temperature (SST) anomaly compared to Pre-Industrial (PI) conditions of up to 1 K was found in the Banda Sea at 125 ka (MIS 5) and 579 ka (MIS 15). The coolest boreal summer SST anomaly down to –2 K at 495 ka (MIS 13) is equally distributed in Indonesian waters. During boreal winter, most of the moderate cooling over large portions of the land and the waters of Indonesia is also associated with local insolation. The most interesting finding in this study, a dipole and tripole precipitation pattern with up to 3.6 mm/day of rainfall anomaly during boreal summer is identified in the western part of the Indonesian waters, Indian Ocean to Banda Sea, and the eastern part of Indonesian waters. The results of this study are expected to be used as basic information to predict the climate in Indonesia for the present and future. This may add to the assessment provided by the IPCC for a better understanding of future climate change in the region, which is a prerequisite for alleviating its impacts.

Zusammenfassung

Wir analysieren die Dynamik indonesischer Gewässer anhand der Ergebnisse von 13 Zeitscheibenexperimenten, die mit dem CCSM3-DGVM-Modell durchgeführt wurden. Die Experimente helfen, globale Klimaschwankungen zwischen und innerhalb der quartären Interglaziale der Marinen Isotopenstufen (MIS) 1, 5, 11, 13 und 15 zu untersuchen. Während des borealen Sommers (Juni-Juli-August-September) können in den meisten Gebieten Indonesiens saisonale Oberflächentemperaturanomalien weitgehend durch lokale Sonneneinstrahlungsanomalien erklärt werden, die durch den astronomischen Antrieb hervorgerufen werden. Allerdings können Klimarückwirkungen für einige Zeitscheiben die Oberflächentemperaturen in und um Indonesien verändern, am stärksten im offenen Ozean in der Nähe des Indischen und Pazifischen Ozeans. Die wärmste Anomalie der Meeresoberflächentemperatur (SST) im borealen Sommer im Vergleich zu vorindustriellen Bedingungen um bis zu 1 K wurde in der Banda-See bei 125 ka (MIS 5) und 579 ka (MIS 15) gefunden. Die kühlste boreale Sommer-

Rima Rachmayani, Matthias Prange, Michael Schulz, Nining Sari Ningsih 2019: Climate Variability in Indonesia from 615 ka to present: First Insights from Low-Resolution Coupled Model Simulations. – DIE ERDE **150** (4): 230-240

DOI:10.12854/erde-2019-428

SST-Anomalie um bis zu -2 K bei 495 ka (MIS 13) ist in indonesischen Gewässern gleichmäßig verteilt. Während des borealen Winters ist die meist moderate Abkühlung über weite Teile des Landes und der Gewässer Indonesiens auch mit der lokalen Sonneneinstrahlung verbunden. Das interessanteste Ergebnis dieser Studie, ein Dipol- und Tripolniederschlagsmuster mit bis zu 3,6 mm/Tag Niederschlagsanomalie während des borealen Sommers, wird im westlichen Teil der indonesischen Gewässer, im Indischen Ozean bis zur Banda-See und im östlichen Teil der indonesischen Gewässer identifiziert. Die Ergebnisse dieser Studie sollen als Basisinformationen dienen, um das Klima in Indonesien für die Gegenwart und Zukunft vorherzusagen. Dies kann die vom IPCC vorgelegte Bewertung für ein besseres Verständnis des zukünftigen Klimawandels in der Region ergänzen, als eine wichtige Voraussetzung für die Abschwächung seiner negativen Auswirkungen.

Keywords Indonesia, surface temperature, precipitation, CCSM3

1. Introduction

Indonesia is located within the Indo-Pacific Warm Pool (IPWP) which is known as the largest area of warm sea-surface temperatures (SST) (*Niedermayer* et al. 2014; *De Deckker* 2016; *Wurster* et al. 2019) with the highest precipitation on the entire Earth (*Niedermayer* et al. 2014; *Wurster* et al. 2019). Thus, the IPWP plays a major role in the global atmospheric circulation and hydrologic cycle, and the IPWP can be considered as the largest source of atmospheric water vapor and latent heat. In Indonesia, marine and terrestrial ecosystems are strongly dependent on the global and regional climate evolution. Today, the Australian–Indonesian monsoon system and the migration of the Intertropical Convergence Zone (ITCZ) passing Indonesia drive the seasonal cycle of the Indonesian climate (*Wyrtki* 1961; *Robertson* et al. 2011; *Kwiatkowski* et al. 2015). Meanwhile, positive climate anomalies like the Indian Ocean Dipole (IOD) (*Saji* et al. 1999) or/and El Niño-Southern Oscillation (ENSO) events (*Rasmusson* and *Carpenter* 1982) have been suggested to contribute to drier conditions and prolonged dry seasons in Indonesia. Interaction among these recurrent modes may generate a very complex climatic system over Indonesia. Along with the strong entanglement of climate phenomena, their influence varies across the region today and may have experienced similar climate features during the past due to, for example, the topography of Indonesia and ocean–atmosphere fluxes, which are mainly imposed by SST variability (*Aldrian* and *Susanto* 2003), partly induced by insolation forcing.

Paleoclimate conditions in Indonesia, particularly during the Holocene, have been studied with proxies by using lacustrine (*Konecky* et al. 2013; *Russell* et al. 2014) and marine sediment cores (e.g. *Linsley* et al. 2010; *Mohtadi* et al. 2011a), corals (*Abram* et al. 2009), and speleothem records (e.g. *Partin* et al. 2007; *Griffiths* et al. 2009; *Griffiths* et al. 2010). Proxy records were collected from marine sediments off Sumatra, western Indonesia, from the Makassar Strait, central Indonesia, and the Java Sea, southern Indonesia (*Partin* et al. 2007; *Griffiths* et al. 2009; *Griffiths* et al. 2010; *Mohtadi* et al. 2011a; *Tierney* et al. 2012; *Ayliffe* et al. 2013; *Konecky* et al. 2013; *Dubois* et al. 2014; *Russell* et al. 2014; *Steinke* et al. 2014; *Kuhnt* et al. 2015). These reconstructions were accomplished for different regions in the Maritime Continent dominated by large-scale climate phenomena such as ENSO or the monsoonal system to a different extent as it is observed today (*Aldrian* and *Susanto* 2003). The reconstructions revealed that the monsoonal system (*Mohtadi* et al. 2011b), ENSO and the IOD (*Abram* et al. 2009; *Niedermeyer* et al. 2014) may have transformed over time.

With regard to the past interglacials, the present Holocene global climate pattern and its natural near future is best comparable with Marine Isotope Stage (MIS) 19 at ~790–760 ka (*Lisiecki* and *Raymo* 2005) based on variations in both annual and seasonal temperatures; the MIS 11 (~424–373 ka) climate pattern is close to the Holocene climate pattern when the impact of insolation alone is considered, and warm climates in MIS 5 (~130–70 ka) and MIS 9 (~337–299 ka) make the closest analogues to the future human-induced warm climate (*Yin* and *Berger* 2015). This insolation-induced climate pattern may occur again in the future, resulting in a similar climate pattern as was observed in the past. Hence, investigating the mechanisms that have been essential in the past may help in understanding the future (*Doe* 1983; *Hay* et al. 1997). However, anthropogenic influence, as observed today, may modify the climatic pattern on Earth also in the future, in particular in Indonesia. Facing global climate change, the average temperature changes are projected to rise by 3°C in Indonesia by 2100 relative to the average temperature of 1990; sea level height may rise up to 140 cm by 2080 (*BAPPENAS* 2011). Hence, future environmental changes may lead to intensified monsoonal precipitation (*Jourdain* et al. 2013; *Pachauri* et al. 2014), shifted El Niño conditions (*Collins* 2005), and altered agricultural activities in Indonesia (*Naylor* et al. 2007) which will result in a hampered food security and nutrients supply to local people.

Effects of environmental changes can be determined by using climate simulations. Thus, climate models are of pivotal importance to simulate future climate scenarios and to identify dynamics and forcing mechanisms of climate phenomena in response to orbital forcing and greenhouse gases (GHG) in order to understand how climatic phenomena are interconnected in the past, present, and future. Moreover, neither one of interglacial climate simulations focusing on Indonesia have been performed in previous studies using Earth system models of intermediate complexity or fully coupled atmosphere–ocean general circulation models. In addition, earlier interglacial periods have attained much less attention by climate modellers.

The aim of this study is to investigate the climate evolution in Indonesia on long time-scales from 615 ka to present related to orbital forcing and GHG concentrations by using the Community Climate System Model version 3 (CCSM3) including a dynamic global vegetation model (DGVM) following the previous work of *Rachmayani* et al. (2016). The results of our study are expected to be useful as basic information to understand the climate, especially in Indonesia for the present and future. Moreover, this research indirectly plays a role in alleviating the impacts of climate change (disaster mitigation) which is in line with the goals of the Intergovernmental Panel on Climate Change (IPCC).

2. Experimental setup

2.1 Model

To investigate the Indonesian climate dynamics, we conducted low-resolution T31 version (*Yeager* et al. 2006) simulations with the CCSM3 coupled general circulation model (CGCM) comprised of four components representing (1) atmosphere, (2) ocean, (3) sea ice, and (4) land surface (*Collins* et al. 2006). In this T31 version, the horizontal resolution of the atmosphere and land components is 3.75° with 26 vertical layers in the atmosphere. The ocean grid consists of 25 levels in the vertical and nominal horizontal resolution of 3°. Improvements of land hydrology parameterizations (*Oleson* et al. 2008) were carried out as in previous studies (e.g. *Rachmayani* et al. 2015; *Rachmayani* et al. 2016). The land model consists of components for biogeophysics, biogeochemistry, the hydrological cycle and a DGVM based on the Lund–Potsdam–Jena (LPJ) model (*Sitch* et al. 2003; *Levis* et al. 2004; *Bonan* and *Levis* 2006).

2.2 Experimental setup

A control simulation of standard Pre-industria (PI) was performed following PMIP (Paleoclimate Modelling Intercomparison Project) guidelines with regard to the forcing (see e.g. *Braconnot* et al. 2007). The astronomical parameters of PI in 1950 AD, atmospheric trace gas concentrations from the 18th century (*Table 1*) as well as PI distributions of atmospheric ozone, sulfate aerosols, and carbonaceous aerosols (*Otto-Bliesner* et al. 2006) were considered along with the solar constant set to 1365 Wm^{-2}. To begin the simulation, 1000 years of the PI control run were integrated starting from modern initial conditions, except for the vegetation which started from bare soil. Branching off from year 600 of the PI spin-up run, 13 interglacial time-slice experiments were executed, each running for 400 years. *Table 1* comprises the appointed time slices according to the astronomical parameters (*Berger* 1978) and GHG concentrations, whereas ice-sheet configuration, ozone distribution, sulfate aerosols, carbonaceous aerosols, and solar constant were prescribed as in the PI control run. Here, we used the mean of the last 100 simulation years of every experiment for analysis. The selection of interglacial time slices was discussed in *Rachmayani* et al. (2016). The time slices were categorized into three groups according to the insolation patterns which diverge in their seasonal distribution of incoming energy (*Rachmayani* et al. 2016). Group I is composed of 6 and 9 ka (MIS 1), 125 ka (MIS 5), 405 and 416 ka (MIS 11), 504 ka (MIS 13), and 579 ka (MIS 15) time slices which reflected high northern hemisphere summer insolation anomaly. Group II consists of dates 115 ka (MIS 5), 495 and 516 ka (MIS 13), and 609 ka (MIS 15) which showed anomalies with low boreal summer insolation. Group III includes dates 394 and 615 ka, which are characterized by changes in the sign of the northern hemisphere insolation anomalies from spring to summer (*Rachmayani* et al. 2016).

Table 1 Atmospheric greenhouse gas (GHG) concentration used in the interglacial experiments. Source: Rachmayani et al. (2016)

Group classification*	Stage	Time slice (ka)	CO_2 (ppmv)	CH_4 (ppbv)	N_2O (ppbv)
Control run	MIS 1	0	280	760	270
I		6	280	650	270
I		9	265	680	260
II	MIS 5	115	273	472	251
I		125	276	640	263
III	MIS 11	394	275	550	275
I		405	280	660	285
I		416	275	620	270
II	MIS 13	495	240	487	249
I		504	240	525	278
II		516	250	500	285
I	MIS 15	579	252	618	266
II		609	259	583	274
III		615	253	617	274

* Group I:
High northern hemisphere summer insolation anomaly

Group II:
Low northern hemisphere summer insolation anomaly

Group III: Changes in the sign of the northern hemisphere spring to summer insolation anomaly

CCSM3 simulations were performed on the SGI Altix supercomputer of the Norddeutscher Verbund für Hoch- und Höchstleistungsrechnen (HLRN), Hannover, Germany, through the Priority Research Programme INTERDYNAMIC (SPP 1266). The compute node details comprise of 960 eight-core central processing unit (CPU) sockets sharing 48 GB of random access memory (RAM). The total memory used in Hannover was 45 TB with 7680 internal service nodes. The computation time to run one single 1000 years paleoclimate simulation by using the T31 low resolution of CCSM3-DVGM was ~5–6 days.

3. Results and discussion

3.1 Surface temperature

Figure 1 displays surface temperature anomalies during boreal summer (June-July-August-September) in Indonesia for each time slice. *Rachmayani* et al. (2016) divided the time slices into three groups based on the insolation (see Section 2.2). As in *Rachmayani* et al. (2016), in most of Indonesia, seasonal surface temperature anomalies can largely be interpreted by local insolation anomalies induced by the astronomical forcing (not shown; see *Fig. 2* in *Rachmayani* et al. 2016). However, for some time slices such as at mid-Holocene (6 ka), at 394 ka and at 615 ka, climate feedbacks may have modified the surface temperature response in Indonesia, most pronounced in open waters close to the Indian Ocean and the Pacific Ocean.

The warmest SST anomaly of up to 1 K is in the Banda Sea at 125 ka (MIS 5) and 579 ka (MIS 15) compared to PI conditions. The coolest SST anomaly of down to -2 K, equally distributed in Indonesian waters, is captured at 495 ka. During boreal winter (December, January, February), the local insolation anomaly contributes to the most of moderate cooling over large portions of land and waters of Indonesia at 6, 9, 125, 394, 405, 416, 504, and 579 ka with maximum cooling at 579 ka of down to -2.5 K, compared to PI (*Fig. 2*). A warmer winter surface temperature is observed at 115, 609, and 615 ka with an anomaly of about 0.5 K compared to PI. Meanwhile, surface temperature anomaly at 516 ka is small compared to PI.

During boreal summer in the Holocene, an abrupt warming of surface temperature can be associated with a weakening of the Asian summer monsoon and a more southerly displaced ITCZ, probably reflecting negative IOD-like mean states (*Abram* et al. 2009) in the Indian Ocean. By contrast, during winter, in line with *Linsley* et al. (2010), SST experiences a cooling during the early Holocene (9 ka) related to seasonal migration to a more northerly position of the ITCZ and associated monsoonal precipitation as discussed in *Fan* et al. (2013) and a strong Asian summer monsoon corresponding to a more positive IOD-like mean state in the Indian Ocean (*Abram* et al. 2009). Some feedback mechanism such as an eastward shift of the Western Pacific Warm Pool (WPWP) or intensified ENSO may contribute to the cooling along with an upward mixing of cold subsurface waters within the Indonesian waters as suggested by *Rosenthal* et al.

(2013). Here, we hypothesize that a similar phenomenon happened at 125, 416, 504, and 579 ka associated with a stronger change of ITCZ and larger shifted WPWP.

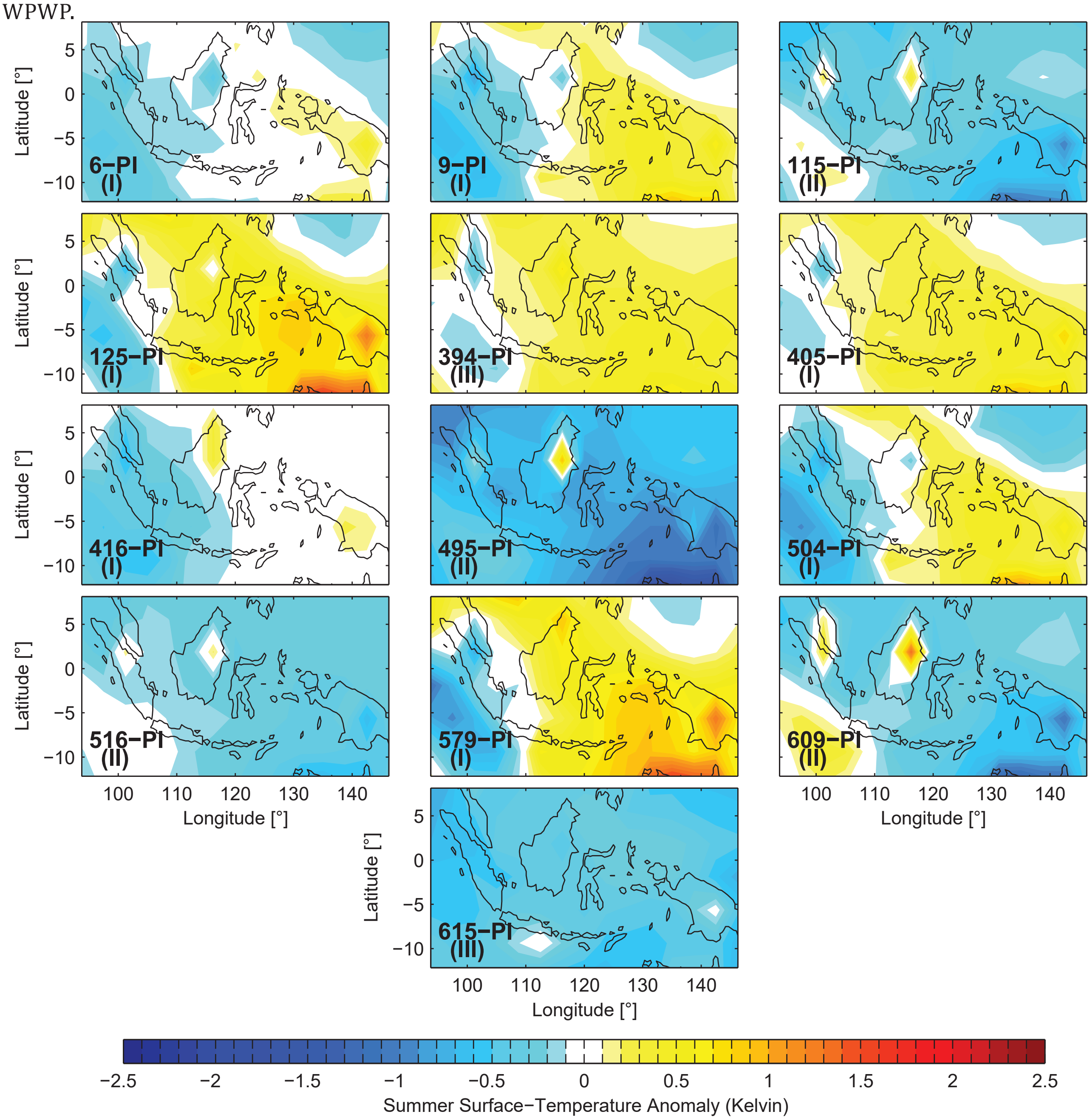

Fig. 1 *Boreal summer (June–September) surface temperature anomalies (relative to Pre-industrial times, PI) for the different interglacial time slices (numbers are given in 1000 yrs). Classification as Groups I, II, and III according to Table 1 is indicated. Source: own elaboration*

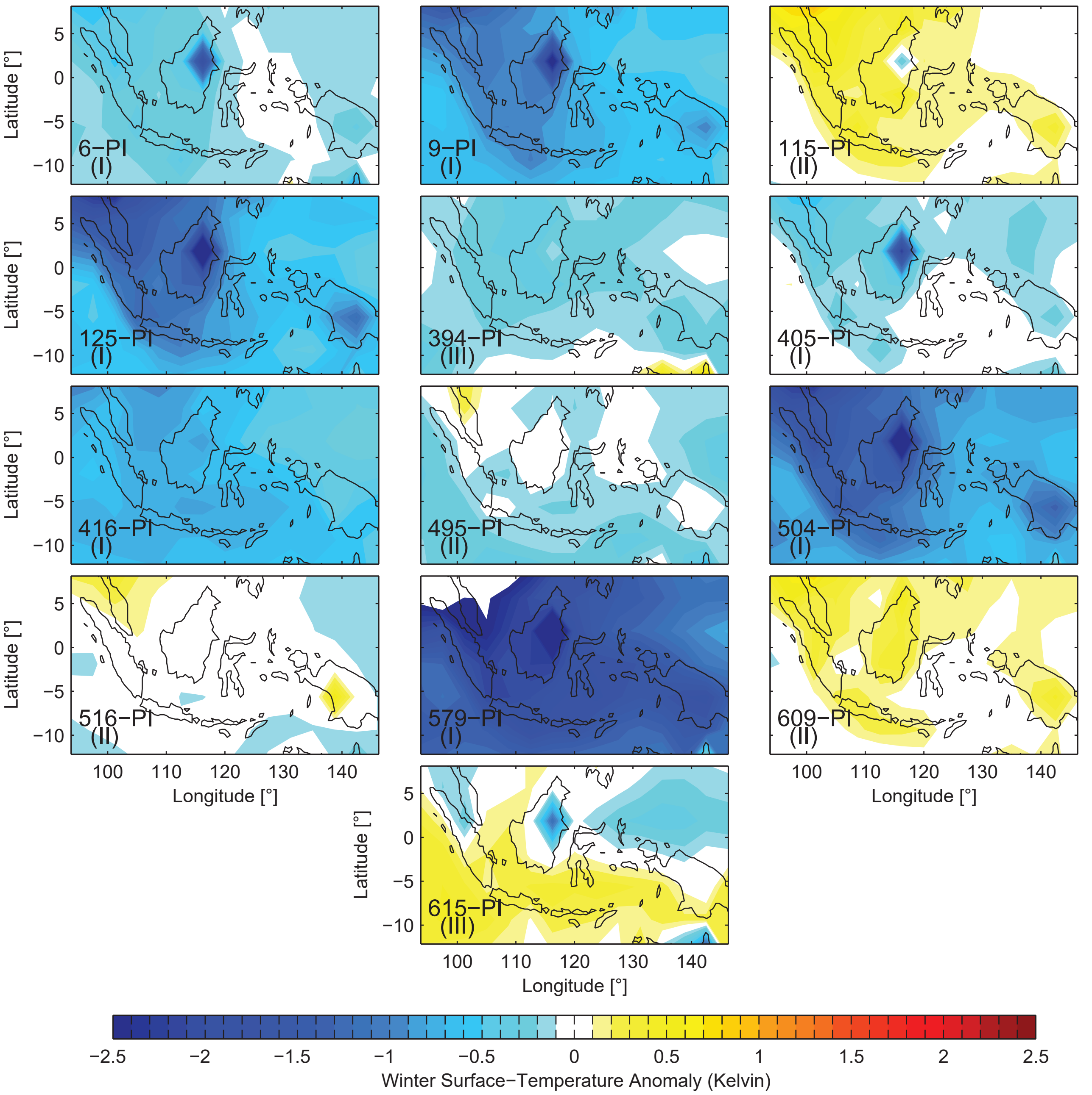

Fig. 2 *As in Figure 1, but for boreal winter (December–February). Note: White area in the panel for 579-PI indicates values below -2.5 K. Source: own elaboration*

3.2 Precipitation

Boreal summer (JJAS) precipitation over Indonesia is presented in *Figure 3*. It illustrates intensified rainfall in the northwest and eastern region of Indonesia in Group I as a response to the high summer insolation and Group III due to internal feedbacks. It reaches the maximum rainfall anomaly at about 3.6 mm/day at 125, 504 and 579 ka compared to PI. This is associated with low precession values (*Rachmayani* et al. 2016). In addition, these two regions experienced a wet phase as an extension of the monsoon belt from northern Africa to India, via the Arabian Peninsula as discussed in *Rachmayani* et al. (2016). Contrastingly, the two regions evidence dry conditions in Group II with low boreal summer insolation due to a precession maximum (*Rachmayani* et al. 2016). The most interesting part in this study, a dipole and tripole precipitation pattern is captured in the western part of Indonesian waters, Indian Ocean to Banda Sea and eastern part of Indonesian waters during boreal summer.

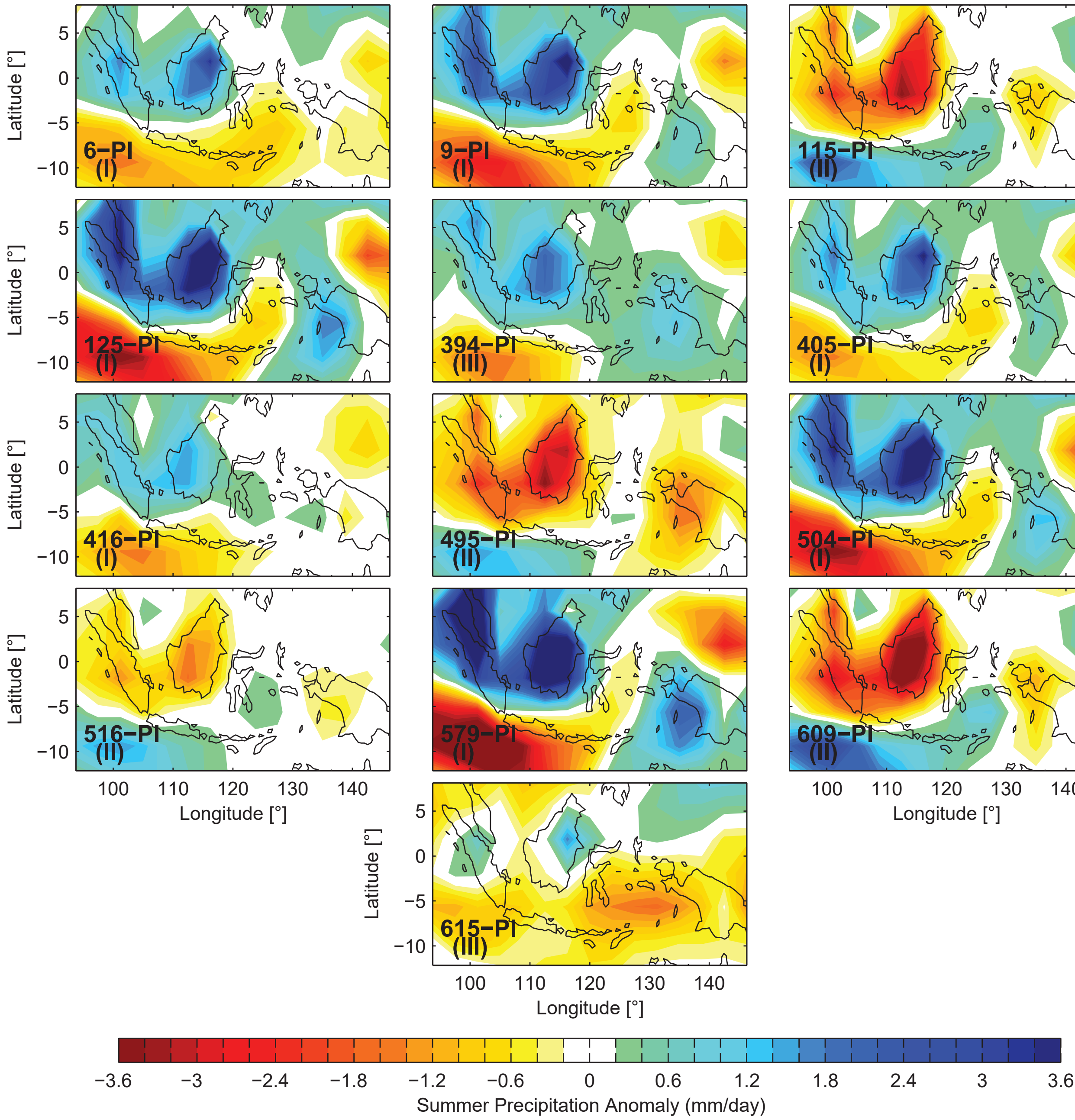

Fig. 3 *As in Figure 1, but for precipitation. Source: own elaboration*

As investigated in many studies, the tropical climate of Indonesia is controlled by the ITCZ and the Australian–Indonesian monsoon system on a seasonal scale (*Robertson* et al. 2011), and ENSO (*Dai* and *Wigley* 2000) and IOD (*Saji* et al. 1999; *Webster* et al. 1999) on an interannual scale. The position of the ITCZ is strongly controlled by insolation (*Wanner* et al. 2008) and oceanic energy transport (*Broccoli* et al. 2006), which affects the interhemisphere temperature contrast (*Chiang* and *Friedman* 2012). Moreover, it is tightly coupled to the large-scale phenomena like the monsoonal system (*Gadgil* 2003; *Wang* 2009) and ENSO (*Philander* 1985). The position of the ITCZ being located farther south during El Niño and farther north during La Niña (*Philander* 1985; *Schneider* et al. 2014). The intensified rainfall in the northwest region of Indonesia and eastern part of Indonesia is also associated with the WPWP. It is characterized by the warmest SST (in the Pacific Ocean) and it is known as the largest heat reservoir on Earth providing water vapor and latent heat to the atmosphere (*Chen* et al. 2004; *Cravatte* et al. 2009) influencing the tropical climate system such as in Indonesia at dates in Group I and Group III. By contrast in Group II, the shifted WPWP causes a reduction of rainfall in Indonesia. The monsoonal system (wind strength, precipitation intensity and the onset of the rainy season) over Indonesia is influenced by interannual climate from the Indian and Pacific Oceans. Monsoon strength changes and land–ocean movements of rainfall during the interglacials are likely affected by the orbital parameters and insolation gradient resulting in changes in the meridional temperature gradient (e.g. *Mohtadi* et al. 2016).

4. Conclusions

Since past climate variability in Indonesia is mostly studied by using proxies, this study is an attempt to investigate the climate variability in Indonesia by using the fully coupled CCSM3-DGVM climate model. Thirteen interglacial time slices of the Indonesia region were analyzed with respect to climate variability between and within Quaternary interglacials. We focused on the local and regional climate feedbacks which may modify the climate variability from the external forcing of insolation and GHG concentrations. The climate feedbacks that play a role in contributing to the rainfall pattern in Indonesia involve the position of the ITCZ, the Australian–Indonesian monsoonal system, and the WPWP. Compared to the past interglacials, the present global climate pattern and its near future has similarities with the MIS 19, MIS 11, MIS 9 and MIS 5 climate patterns. Periodic variations in the distribution of insolation strongly influenced climatic patterns on Earth in the past, present and will influence them in the future. Facing global climate change, the average temperature is projected to rise by 3°C in Indonesia by 2100 relative to the average temperature in 1990, while sea level height may rise up to 140 cm by 2080. Hence, future environmental changes may lead to intensified monsoonal precipitation, shifted El Niño conditions, and altered agricultural activities in Indonesia, resulting in hampered food security and nutrients supply to local people. With increasing computer power, long-term simulations of interglacial climates will become more common. Simulations with a higher spatial resolution of the model will help to develop a significantly deeper understanding of the complexity of interglacial climate variability in Indonesia during the past to investigate climate variation of various regions in Indonesia. In addition, simulations with various GHG concentration trajectories by using the RCPs (Representative Concentration Pathways) adopted by the IPCC are needed to project the future climate dynamics in Indonesia.

Acknowledgments

R. R. acknowledges the funding support from Deutsche Forschungsgemeinschaft (DFG) within the COmparison of INterglacials (COIN) project, grant number PR 1050/1-2 during her research period in Germany. Moreover, we acknowledge support received from the Institut Teknologi Bandung via their Program Penelitian dan Pengabdiaan Masyarakat (P3MI) and from Ministry of Research, Technology, and Higher Education (Kemenristekdikti) of the Republic of Indonesia under the research program of Penelitian Dasar Unggulan Perguruan Tinggi (PDUPT) 2019.

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