The spatio-temporal evolution of the Asian monsoon climate in the Late Miocene and its causes: A regional climate model study

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Cover photo: “Monsoon boundary” over the Loess Plateau, Qin’an, China.

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Abstract

Late Miocene (11-5 Ma) is a crucial period for Asian monsoon evolution. Significant changes of both the India and East Asian monsoons in this period have been documented in geological records. However, the relative strength and the spatio-temporal structure of the Asian monsoon climate in this period remain ambiguous, and the contributions of different forcings, such as global cooling and the uplift of the Tibetan Plateau (TP), to the monsoon changes in this period are still controversial.

In this study, a high-resolution regional climate model COSMO-CLM driven by a fully coupled atmosphere-ocean general circulation model ECHAM5/MPI-OM is employed to investigate the monsoon climate changes in the Late Miocene. We firstly performed a regional climate simulation for the early Late Miocene (i.e. the Tortonian) (11-7 Ma). It shows a generally weaker (stronger) summer (winter) monsoon circulation over East Asia and India in the Tortonian than today, which agrees with its global driving model. The regional difference of monsoon rainfall changes between southern and northern China (India), however, is highlighted by the regional model experiment, due to its better characterization of small-scale convective activity in the monsoon regions. It is indicated that a modern-like monsoonal climate may have existed in southern China and India in the Tortonian, while the monsoonal climate in northern China and northwestern India may have not fully developed at that time.

To further evaluate the role of mountain uplift in driving the Asian monsoon changes in the Late Miocene, we performed sensitivity experiments with different orographic configurations in our regional model to depict the diachronous growth of different part of the Asian orography (i.e. regional mountain uplift) at that time. The results show that different from the effect of the uplift of the entire Asian orography (i.e. bulk mountain uplift), which strengthens both the Indian and East Asian summer monsoons (ISM and EASM), the regional mountain uplift can lead to an asynchronous development of the ISM and EASM. While the ISM is primarily intensified by the presence of the southern TP (Zagros Mountains) due to its thermal insulation (mechanical blocking) effect, the EASM is mainly strengthened by the presence of the central and northern TP and the Tianshan Mountains owing to their strong diabatic heating effect. It is found that the asynchronous development of the ISM and EASM during the Middle Miocene (15 Ma), the Late Miocene (8 Ma) and Pliocene (4 Ma) may have been closely associated with the uplift of the southern TP, northern TP and the Zagros Mountains, respectively.

In addition to the mean monsoon climate, the interannual variability of the ISM in the Late Miocene is also explored by analyzing the 110-year integration of global and regional experiments for the Tortonian. It reveals that the interannual variability of the ISM in the Late Miocene may have been as strong as, or even stronger than at present. This can be attributed to the strong El Niño–Southern Oscillation (ENSO)
at that time. In addition, the extratropical influence, such as summer North Atlantic Oscillation, on the ISM might have been weak. This may have facilitated a stronger ENSO-ISM teleconnection in the Late Miocene.

We note that although the Late Miocene Asian monsoon climate is strongly influenced by regional tectonic changes (e.g. the lower northern TP), it still exhibits great resemblance to that projected under the future global warming conditions. Better knowledge and model simulations on the Late Miocene monsoon climate may provide useful constraints on the prediction of the future Asian monsoon changes.

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“學而不思則罔, 思而不學則殆”
—孔子

“Without thinking, learning will be confused. Without learning, thinking will be dubious.”
— Confucius
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List of original publications

This thesis is based on the following peer-reviewed publications:

Paper I:  **Tang H.,** Micheels A., Eronen J. T., Fortelius M. 2011. Regional climate model experiments to investigate the Asian monsoon in the Late Miocene. Climate of the Past 7: 847-68


The publications are referred to in the text by their roman numerals.

Author’s contributions

Paper I:  The study was designed by HT, AM, JTE and MF. AM and HT conducted the global and regional model experiments, respectively. HT analyzed the results and prepared the manuscript with the contributions from the other coauthors.

Paper II:  The study was designed by HT, AM, JTE and MF. HT conducted the regional model experiments, analyzed the results and prepared the manuscript with the contributions from the other coauthors.

Paper III:  The study was designed by HT, AM, JTE and BA. AM and HT conducted global and regional model simulations, respectively. HT analyzed the results and prepared the manuscript with the contributions from the other coauthors.
Abbreviations

AIMR, EIMR, IMI, MH, WY, SWI, SWII  Summer monsoon indices (see Table 2 in Synopsis)
CGCM  Coupled general circulation model
CTRL  Regional model present-day climate control run
DJF  December-January-February
EASM  East Asian summer monsoon
EAWM  East Asian winter monsoon
ENSO  El Niño-Southern Oscillation
GCMs  General circulation models
GCTRL  Global model present-day climate control run
GTORT  Global model Tortonian climate run
ISM  Indian summer monsoon
ITCZ  Inter-tropical convergence zone
(M)JJA(S)  (May)-June-July-August-(September)
Ma  Million years ago
MO, WJ  Winter monsoon indices (see Table 2 in Synopsis)
Mxx  Mountain uplift experiments (See Table 1 and 3 in Paper II)
NAO  North Atlantic Oscillation
PDTORT  Regional model experiment with Tortonian physical boundary conditions
RCMs  Regional climate models
TORT  Regional model Tortonian climate run
TORTPD  Regional model experiment with Tortonian global forcing.
TP  Tibetan Plateau
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1. Introduction

The Asian monsoon is one of the most active components in the modern global climate system. The presence of the Asian monsoon can be traced back to early Cenozoic (Huber & Goldner 2012, Quan et al 2012, Z.S.Zhang et al 2012), and a multi-phase development of the mean Asian monsoon climate has been documented by various geological proxies (An 2000, Clift et al 2008, Wang et al 2005). Among those periods, the Late Miocene (11.6–5.3 Ma) is widely regarded as a crucial period in the Asian monsoon history (An et al 2001, Molnar et al 2010). An inception or marked strengthening of the monsoon system in both South Asia (Kroon et al 1991, Quade et al 1989, Sanyal & Sinha 2010) and East Asia (For telius et al 2002, Rea et al 1998) in this period have been inferred from various proxy records, which witnessed the formation of modern Asian monsoon climate.

Large uncertainties, however, still remain over the spatial and temporal evolution of the Asian monsoon in the Late Miocene. For instance, there has been controversy over the relative strength of the Asian summer and winter monsoon in the Late Miocene. While the early evidence suggests weaker-than-present summer (Chen et al 2003, Fortelius et al 2002, Huang et al 2007) and winter (Rea et al 1998, Wan et al 2007, Wang et al 2003) monsoons, recent studies claim stronger-than-present ones (Dettman et al 2001, Jiang & Ding 2009, Steinke et al 2010, Q.Q.Zhang et al 2012). There is also debate over the distribution of monsoon climate in the Late Miocene, particularly over northern China. A similar-to-present monsoon climate distribution since the Early Miocene has been documented by Sun and Wang (2005). But recent climate reconstructions based on plant and mammal fossil records reveal that the present-day distribution of monsoon climate over northern China might have not established until the Late Miocene (L.P.Liu et al 2009, Y.S.Liu et al 2011). The complex spatial structure is an inherent feature of the modern Asian monsoon systems (Ding 1992, Gadgil 2003). Evidence indicating regional heterogeneity of the Asian monsoon climate and the corresponding vegetation changes in the late Miocene has recently emerged (Eronen et al 2010a, Fortelius et al 2002, Passey et al 2009, Sanyal et al 2010). In particular, the accumulating spatially resolved quantitative climate data sets for the Neogene in Eurasia based on plant fossils (NECLIME database: http://www.neclime.de/) and mammal fossils (NOW database: http://www.helsinki.fi/science/now/) have provided valuable information on the spatio-temporal evolution of the Asian monsoon in the Late Miocene. But our understandings on these spatial patterns are very limited.

Various mechanisms have been advanced to explain the Asian monsoon changes in the Late Miocene. The regional tectonics, such as the uplift of the Tibetan Plateau (TP) (An et al 2001, Kutzbach et al 1993, Liu & Yin 2002, Ruddiman & Kutzbach 1989) and the retreat of the Paratethys (to a less extent) (Fluteau et al 1999, Ramstein et al 1997), have been considered as the main drivers for the intensification of both the winter and summer monsoons in the Late Miocene in earlier studies. Recent studies, however, argue that the Asian monsoon strength in the Neogene and the previous periods might have been largely modulated by global climate changes (Clift et al 2008, Dupont-Nivet et al 2007, Ge et al 2012). The global cooling trend and ice sheet growth in the polar regions from the Early-Middle Miocene to Quaternary might have led to the decline of the Asian summer monsoon during the Late Miocene (Clift et al 2008, Jiang & Ding 2009). So far, the relative importance of these mechanisms in determining the Late Mio-
cene Asian monsoon climate is still ambiguous. Climate models have been widely used to simulate and understand the Late Miocene climate changes (Bradshaw et al 2012, Dutton & Barron 1997, Knorr et al 2011, Lunt et al 2008, Micheels et al 2011, Micheels et al 2007, Steppuhn et al 2007, Steppuhn et al 2006). All these studies use global climate models, i.e. general circulation models (GCMs), with coarse spatial resolution, thus lack important regional details over the Asian monsoon area. Because of the coarse resolution, they are also unable to capture the climatic influence of small-scale topography, such as the southern TP and Himalayas, which may have been essential for the Asian monsoon in the Late Miocene (Boos & Kuang 2010, Harris 2006). Compared to GCMs, regional climate models (RCMs) with high spatial resolution can resolve small-scale physical and dynamical processes, therefore performing better than GCMs in capturing the spatial patterns and the magnitude of the Asian monsoon (Dobler & Ahrens 2010, Gao et al 2008, 2006). They have been applied to investigating the Asian monsoon in the mid-Holocene and the Last Glacial Maximum, and demonstrated better agreement with relevant geological evidence (Ju et al 2007, Zheng et al 2004).

In this study, we employed the regional climate model COSMO-CLM driven by a fully-coupled atmosphere-ocean general circulation model (CGCM) ECHAM5/MPI-OM (Micheels et al., 2011) to simulate the Asian monsoon climate in the early Late Miocene, i.e. Tortonian (11–7 Ma). We first analysed the mean Asian monsoon climate simulated by the regional model to demonstrate the regional pattern of the monsoon climate in the Tortonian and the adding values of the regional model results compared to that of the global model. We then performed sensitivity experiments with the regional model to examine the relative importance of the regional tectonics and global climate in shaping the Late Miocene Asian monsoon climate. Special attention has been paid to the influence of regional mountain uplift on the Asian summer monsoons, which is the most controversial part in our understanding of the Late Miocene monsoon evolution (Molnar et al 2010). Furthermore, we took advantage of our global and regional model experiments to investigate the interannual variability of the Asian summer monsoon (mainly the Indian summer monsoon) in the Late Miocene. This was an important part of the Asian monsoon evolution in this period, but has been seldom touched in previous studies. Finally, a comparison of our modelling results with the quantitative climate reconstructions based on plant and mammal fossil data was inspected to recognize the problems with both the model and proxy approaches in representing the Late Miocene monsoon climate. Since the magnitude of the global warming in the Miocene and Pliocene are likely to be comparable to that in the future (Haywood et al 2009, Kutzbach & Behling 2004, Lunt et al 2008, Utescher et al 2011), we are hoping that our improved understanding of the Asian monsoon climate in the Late Miocene may provide useful insights to its changes in the future.

2. Material and methods

2.1. Regional climate model

In this study, we use the regional climate model COSMO-CLM, which is the climate mode of a non-hydrostatic regional weather prediction model COSMO (Consortium for Small-scale Modelling) (available at http://www.clim-community.eu). The model uses a rotated geographical coordinate system. Its vertical domain is represented by a terrain-following hybrid coordinate system ($\eta$ coordinate). In all our studies, we adopted the model version COSMO-CLM2.4.11.
and chose the leapfrog numerics, the Tiedtke convection scheme (Tiedtke 1989), the radiation transfer scheme based on Ritter and Geleyn (1992), the prognostic turbulent kinetic energy closure (Raschendorfer 2001) and the TERRA-ML multi-layer soil-vegetation-atmosphere-transfer model (Schrodin & Heise 2002). Our regional model domain covers the Asian monsoon area (0 to 60 °N and 50 to 140 °E) with a spatial resolution of 1°×1° on the rotated model grid (the north pole is at 60 °N, 80 °W) and 20 vertical levels. We do not use higher spatial resolution in the regional model, because of the large uncertainties on smaller-scale regional boundary conditions in the Late Miocene and the lack of high spatial resolution proxy data to validate the model results. We find the 1°×1° model resolution satisfying for our research purpose (see Sect. 2.5). It also requires less computer hours to run the model. This would allow us to perform more sensitivity experiments to better understand the mechanisms of monsoon changes in the Late Miocene, which is a major interest of this study.

2.2. Initial and lateral boundary forcing

As the initial and lateral boundary forcing for our regional model experiments, we use 6-hourly output from a CGCM ECHAM5/MPI-OM (Eronen et al 2009, Micheels et al 2011). ECHAM5/MPI-OM has been regarded as one of the state-of-the-art CGCMs that represent the Asian monsoon mean and interannual variability relatively well compared to other CGCMs (Hori & Ueda 2006, Kripalani et al 2007a, Kripalani et al 2007b). In this study, the resolution of the spectral atmosphere general circulation model ECHAM5 is T31 (about 3.75°×3.75°) with 19 terrain-following vertical layers. The ocean circulation model MPI-OM uses an Arakawa C-grid with an approximate resolution of 3°×3° and the vertical domain is represented by 40 unevenly spaced levels.

For our regional model experiments, two global model simulations have been used: one is a present-day control run (referred to as GCCTRL), and the other is a Tortonian run (referred to as GTORT). In GTORT, the global orography is broadly reduced. Particularly, the overall elevation of TP is about 70% of its present-day height. The vegetation in GTORT is characterized by less desert and more forest cover than today. Boreal forest is more widespread in high latitudes in GTORT where tundra dominates at present. The present-day ice sheet over Greenland is also absent in GTORT. The land-sea distribution in GTORT is almost the same as today, except that the Paratethys is included and the Panama and Indonesian Seaways are open. Atmospheric CO₂ in GTORT is 360 ppm, which is the same as in GCCTRL. The orbital configuration is also kept as the present-day situation because we aim at representing the Tortonian as an average over 4 million years that contains many orbital cycles. More background on the setup of GCCTRL and GTORT can be found in Micheels et al (2011) and the papers included in this thesis. Both global model runs were integrated over 2600 yr to achieve their dynamic equilibria in terms of global mean temperature and sea ice volume. Only the last 110-year model integrations were used to drive our regional model.

Using the set of adapted Late Miocene boundary conditions, GTORT demonstrates generally warmer (+1.5 °C) conditions compared to GCCTRL (Micheels et al 2011). Global mean precipitation increases by +43 mm year⁻¹ in GTORT due to the higher moisture loading in the atmosphere as a result of the higher surface temperature and greater evaporation over the ocean. Micheels et al (2011) compared the GTORT model results with terrestrial climate proxy data, and showed that GTORT agrees fairly well with proxy data in both the mean an-
nual temperature and precipitation. In general, both GCTRL and GTORT exhibit state-of-the-art performance compared to other present-day and Miocene global model experiments. Therefore, we can use these experiments to force our regional model to simulate the Asian climate in the present-day and the Tortonian.

2.3. Regional model experiments

2.3.1. Present-day and Tortonian runs

All the regional model experiments performed in this study are summarized in Table 1. Two regional model experiments were conducted first to simulate the present-day and Tortonian climate.

For the present-day control run (CTRL), we use the present-day lateral boundary forcing from GCTRL and the present-day physical boundary conditions generated from WEB-PEP (version 0.74) (Smiatek et al. 2008). For the Tortonian run (TORT), we use the Tortonian lateral boundary forcing from GTORT. The physical boundary conditions are modified to better represent the Tortonian conditions.

As to palaeorography, the high spatial resolution of the regional model allows us to modify the topography in more detail compared to the global model. Evidence has shown that the southern TP may have attained its present-day height before the Tortonian (Coleman & Hodges 1995, Polissar et al 2009, Rowley et al 2001, Spicer et al 2003). The elevation of the central and southeastern TP may have also approximated its present-day height in the Tortonin (Blisniuk et al 2001, Liu-Zeng et al 2008, Quade et al 2011, Rowley & Currie 2006, Song et al 2010b). In contrast, the northern TP was mainly uplifted after the Tortonian, and its altitude may have been much lower in the Tortonian than at present (Wang et al 2008b, D.W.Zheng et al 2006, H.B.Zheng et al 2000). To be consistent with these lines of evidence, we kept the elevations of the Himalayas and southern TP at their present-day heights, but reduced the central and southeastern TP to 80% and the northern TP to 30% of their present-day heights in TORT (Fig. 3 in Paper I). The general decrease of elevation from the south to the north of the TP agrees with the stepwise growth of the TP, such as that suggested by Tapponnier et al (2001). For the other topographic relief in our model domain, there are limited palaeoaltimetry studies. Most of the regions have undergone both pre-Tortonian and post-Tortonian surface uplifts, such as the Tianshan Mountains (Charreau et al 2009) and the Zagros mountains (Lacombe et al 2006). Therefore, we simply modified the elevations to 70–90% of their present-day heights in TORT. The roughness length of orography was also reduced proportionally to its surface elevation in TORT.

Table 1. Summary of regional climate model experiments in this study.

<table>
<thead>
<tr>
<th>Experiments</th>
<th>Studies</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Time slice simulations</strong></td>
<td></td>
</tr>
<tr>
<td>CTRL, TORT</td>
<td>The Asian monsoon changes in the Late Miocene: mean state (Paper I); interannual variability (Paper III).</td>
</tr>
<tr>
<td><strong>Sensitivity experiments</strong></td>
<td></td>
</tr>
<tr>
<td>TORTPD, PDTORT</td>
<td>Effect of the regional tectonics and global climate on the Asian monsoon changes in the Tortonian (Paper I).</td>
</tr>
<tr>
<td>M00, MsTibet, McTibet, MnTibet, MZagros, M100, M50</td>
<td>Effect of regional (bulk) mountain uplift on the Asian monsoon changes (Paper II).</td>
</tr>
<tr>
<td>MsTibetNoH, McTibetNoH, MnTibetNoH, MZagrosNoH</td>
<td>Effect of diabatic heating over the mountains on the Asian summer monsoon changes (Paper II).</td>
</tr>
</tbody>
</table>
The modification of vegetation in TORT is largely consistent with that in GTORT. In addition, some regional details of the vegetation were added according to available palaeobotanical data. As illustrated in Fig. 3 in Paper I, we replaced the large desert area in Central Asia by herbaceous vegetation in TORT (Sun & Zhang 2008, Sun et al 2009, Wolfe 1985). The vegetation on the southern TP was changed to mixed leaf trees (Liu 1996, Tang & Shen 1996). Evidence shows that the vegetation of the northern TP and the Loess Plateau underwent significant changes from forest to steppe in the Tortonian (Dong et al 2006, Ma et al 1998, Wang et al 1999). Thus, we covered these areas with open forest, which acts as a transition zone between the forest in the east and the steppe in the west. The range of broad-leaved evergreen forest and subtropical forest is similar to their present-day potential distribution (Hoorn et al 2000, Li & Zhang 1998, Xia et al 2009, Zheng & Wang 1994), but the temperate deciduous forest reaches farther north (Liu 1998, Shu et al 2008, Wang et al 2001b). The prescribed vegetation in TORT is also in agreement with the reconstruction based on vegetation models (Francois et al 2006, Pound et al 2011).

The land-sea distribution in TORT is the same as present, except the presence of the Paratethys. The extent of the Paratethys resembles that in GTORT. The atmospheric CO₂ concentration is 360 ppm for both CTRL and TORT, which is identical to our global model. Both regional model experiments were integrated for 10 years first, which were used in our Paper I for analysing the mean monsoon state. To investigate the interannual variation of the Asian monsoon, both CTRL and TORT were further integrated for another 100 years, which were used in our Paper III. Since the initial adaptation of the upper-level soil moisture in the regional model requires several months (Fig. S1 in Paper I), the first year integration is left for the model to spin up. Only the last 9-year (109-year) results were analysed in Paper I (Paper III).

### 2.3.2. Sensitivity experiments with Tortonian boundary conditions

To further assess the contribution of regional tectonics (e.g. the lower northern TP) and global climate (e.g. the higher global temperature) to the Asian monsoon changes in the Tortonian, two additional regional model experiments were carried out. The first one is driven by the Tortonian global forcing but uses the present-day regional physical boundary conditions (referred to as TORTPD). The other one is driven by the present-day global forcing but uses the Tortonian regional physical boundary conditions (referred to as PDTORT). By comparing the climatologies of TORTPD (PDTORT) with that of CTRL, the climate responses to Tortonian global climate (regional tectonics) can be estimated. Both experiments were integrated for 10 years and the last 9-year results were used for analysis.

### 2.3.3. Sensitivity experiments with different orography

To evaluate the climatic impact of diachronous growth of different mountain regions (i.e. regional mountain uplift) on the Asian monsoon climate, six sensitivity experiments were carried out in our regional model (see Table 1 in Paper II). In the first experiment, all the present-day orography higher than 250 m is removed, which is referred to as M00 (Fig. 2a in Paper II). Then, the present-day southern TP, central TP and northeastern-southeastern TP are added successively. They are referred to as MsTibet, McTibet and MnTibet, respectively (Fig. 2b-d in Paper II). In the experiment MZagros, the present-day height of the complex mountain chains, including the Zagros to the south and the Elburz to the north in Iran, and the Hindu Kush in Afghani-
stan and Pakistan are present (Fig. 2e in Paper II). In the experiment MTian, the present-day Tianshan and Gobi Altai Mountains to the north of the TP are further added (Fig. 2f in Paper II). The control run with all the present-day orography is referred to as M100 (Fig. 2h in Paper II). Moreover, an experiment with all the orography reduced to 50% of its present-day height was also performed (referred to as M50) (Fig. 2g in Paper II) to compare the effects of regional mountain uplift and the uplift of the whole Asian orography (i.e. bulk mountain uplift). Since the palaeo-altitude of the Asian orography is still poorly quantified, our regional mountain uplift experiments intend not to exactly follow, but to generally mimic its growth in geological periods (see Fig. 1 in Paper II). More discussion on the rationale of our experimental design can be found in the Supplement of Paper II.

In all the orographic sensitivity experiments, only the surface elevation was modified. Different from that in TORT, we did not change the roughness length of the orography according to its surface height. This enables us to examine the pure effect of surface height on the monsoon climate. Compared to the effect of surface height, the changes of roughness length have minor influence on the Asian monsoon in our regional model (figure not shown). Other physical boundary conditions, such as land-sea distribution, vegetation and CO$_2$ concentration (360 ppm), are also the same as at present. All the experiments are driven by the 6-hourly output from both the GCTRL and GTORT. Since GCTRL and GTORT exhibit pronounced difference in the mean circulation pattern and climate state, we expected that such difference in global forcings may affect the climatic response to mountain uplift in our regional model. However, such influence turns out to be small. Therefore, only the outcome driven by GTORT was presented in our paper II. The results under the present-day global forcing were briefly analyzed in the Supplement of Paper II. All the regional model experiments were integrated for 10 years and the last 9-year results were used for analysis.

2.3.4. Sensitivity experiments with reduced surface sensible heating

To assess the contribution of sensible heating over the uplifted mountains to the sum effect of mountain uplift on the Asian monsoon, a set of sensitivity experiments with reduced surface heat flux over the uplifted regions (i.e. MsTibetNoH, McTibetNoH, MnTibetNoH, MZagrosNoH) were performed (Table 1). In these experiments, the model setup is the same as in the corresponding mountain uplift experiments, except that the soil type and vegetation of the uplifted regions were changed to ice. This greatly increases the surface albedo, thus reduces the surface sensible heat flux from the uplifted regions (Table 3 in Paper II). We refer to these experiments as the No-Heating experiments. The differences between the No-Heating experiments and the corresponding mountain uplift experiments reflect the sensitivity of monsoon climate to the surface heating of each uplifted region.

2.4. Monsoon indices

In this study, we have calculated various monsoon indices in order to depict the monsoon strength from different aspects (Table 2). The “strength” of monsoon climate can refer to the strength of summer monsoon, winter monsoon, or seasonality. For the strength of summer monsoon, both the circulation and precipitation indices have been employed to reflect the strength of the Indian summer monsoon (ISM) (e.g. WY, MH, IMI, AIMR, EIMR) and the East Asian summer monsoon (EASM) (e.g. SWI and SWII), respectively (Table 2). The meaning of these indices and the reason for choosing these indices
are described in our Paper I and II. Commonly, the circulation and precipitation indices are closely associated with each other. But they can also be decoupled, indicating opposite changes of the summer monsoon strength. In these cases, we use summer precipitation over the whole India (70-90 °E 10-28 °N) (similar to AIMR) as the primary criteria to define the ISM strength in general. Because precipitation is more relevant for palaeomonsoon studies and more widely recorded by proxies. For the EASM, we use summer precipitation over northern China (105-120 °E 34-42 °N) instead of the whole East China as the primary criteria to define its general strength, owing to the large spatial difference of precipitation changes in this region.

For the strength of winter monsoon, two indices (MO and WJ) were used in this study (Table 2). WJ describes the average strength of the low-level northwesterly wind over East China and the East China Sea, which is the basic feature of the East Asian winter monsoon (EAWM). MO characterizes the pressure gradient between the Siberian High over northern Eurasia and the Aleutian Low over North Pacific, which is a major driver for the northwesterly winter monsoon wind over East Asia. Normally, MO and WJ are closely related with each other. But their changes may also be decoupled. In this case, we use WJ as the primary indicator for the winter monsoon strength.

As to the strength of seasonality, it can be measured by the seasonal reversal of the prevailing surface wind direction and speed, which is in accordance to the original definition of the monsoon climate (Li & Zheng 2003). It can also be depicted by the seasonal difference in temperature and precipitation (Liu & Yin 2002), which is more frequently adopted in palaeomonsoon studies. While the seasonality of wind tends to be dominantly influenced by the winter monsoon strength because of the much stronger surface wind in winter than in summer, the seasonality of precipitation is more affected by the summer

Table 2. Summary of monsoon indices used in this study.

<table>
<thead>
<tr>
<th>Index</th>
<th>Definitions</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Summer Monsoon</strong></td>
<td></td>
</tr>
<tr>
<td>AIMR</td>
<td>JJA-mean precipitation over all of India excluding the four hilly meteorological subdivisions (Parthasarathy et al 1992).</td>
</tr>
<tr>
<td>EIMR</td>
<td>JJAS-mean precipitation over the extended Indian region (70–110 °E, 10–30 °N) (Goswami et al 1999).</td>
</tr>
<tr>
<td>WY</td>
<td>JJA-mean zonal wind difference between the lower troposphere at 850 hPa and the upper troposphere at 200 hPa over South Asia (60–110 °E, 5–20 °N) (Paper I and II) or (60–90 °E, 5–20 °N) (Paper III) (Webster et al 1998).</td>
</tr>
<tr>
<td>MH</td>
<td>JJA-mean meridional wind difference between 850 and 200 hPa over the ISM region (70–110 °E, 10–30 °N) (Goswami et al 1999).</td>
</tr>
<tr>
<td>IMI</td>
<td>JJA-mean zonal wind difference at 850 hPa between southern India (60–80 °E, 5-15 °N) and northern India (70-90 °E and 20-30 °N) (Wang et al 2001a).</td>
</tr>
<tr>
<td>SWI</td>
<td>JJA-mean meridional wind velocity at 850 hPa over East China (110-125 °E, 20–40 °N) (adapted from l_110 in Wang et al (2008a)).</td>
</tr>
<tr>
<td>SWII</td>
<td>JJA-mean meridional wind difference at 850 hPa between southern China (110–125 °E, 20–30 °N) and northern China (110–125 °E, 30–40 °N) (adapted from l_110 in Wang et al (2008a)).</td>
</tr>
<tr>
<td><strong>Winter monsoon</strong></td>
<td></td>
</tr>
<tr>
<td>MO</td>
<td>DJF-mean sea level pressure difference between the locations near Irkutsk in Russia (106.1 °E, 52.9 °N) and Nemuro in Japan (145.0 °E, 43.6 °N) (Sakai &amp; Kawamura 2009).</td>
</tr>
</tbody>
</table>
monsoon strength due to more abundant rainfall in summer. Therefore, in this study, we only focus on the winter and summer monsoon strength. The seasonality of wind (precipitation) can be easily inferred from the strength of winter (summer) monsoon.

2.5. Present-day model validation

The regional climate model COSMO-CLM has been used for present-day simulations and future climate projections in different regions (Jaeger et al 2008, Rockel & Geyer 2008). It has been shown to simulate the modern Asian monsoon circulation and precipitation well compared to other regional climate models (Dobler & Ahrens 2010, Lucas-Picher et al 2011, Rockel & Geyer 2008). However, it is still unclear whether the model can faithfully reproduce the modern Asian climate under our model configurations, and therefore eligible for applying to the Late Miocene simulation. In particular, we use $1^\circ \times 1^\circ$ spatial resolution in our regional model, which is coarser than that commonly used for the regional model (e.g. $0.5^\circ \times 0.5^\circ$ or higher). Whether such relatively coarse resolution of the regional model can adequately simulate the Asian monsoon and show improved performance compared to its global driving model needs to be examined.

Fig. 1 illustrates the differences between the modelled present-day surface temperature (GCTRL and CTRL) and that based on observation (CRU TS 2.10 data set). It shows that our driving global model (i.e. GCTRL) has strong warm bias over the TP and other mountain regions compared to the observation in both annual and seasonal mean temperature (Fig. 1a,c,e). Such warm bias is greatly alleviated in the regional model (CTRL) probably due to its better representation of the topography of the TP and other mountains (Fig. 1b,d,f). The overall improvement of the regional model in simulating surface temperature can be seen from the Taylor diagram in Fig. 3a, which displays that the annual, summer and winter temperatures in the regional model are all closer to observation compared to the
global model. Nonetheless, the regional model (i.e. CTRL) inherits some biases from the driving global model, including the warm bias over Central Asia and the cold bias over northern and northeastern China. It also exhibits stronger cold bias over the Indochina and Indian subcontinent than its driving global model. This may be related to its own model physics.

The precipitation differences between model and observation are illustrated in Fig. 2. The driving global model (GCTRL) displays a wet bias over northern and northeastern China, but a dry bias over northern Indian and Indochina subcontinent throughout a year (Fig. 2a,c,e). Over Central Asia, it shows drier conditions in summer but wetter conditions in winter than the observation. Compared to GCTRL, our regional model (CTRL) exhibits similar spatial pattern of precipitation biases (Fig. 2b,d,f), indicating strong influence of the biases from the global forcing data. It, however, displays smaller wet bias over northern China, but more extensive and intense dry bias over India, Indochina subcontinent and southern China. The overall performance of precipitation simulation is only slightly improved in summer and annual mean in CTRL (Fig. 3b). The most significant improvement of CTRL is in simulating winter precipitation (cf. Fig. 2e and f). Particularly, the spatial variance of winter precipitation over Asia is more realistically reproduced in CTRL (Fig. 3b). In general, the spatial pattern and magnitude of precipitation biases in CTRL are comparable to those shown by Rockel and Geyer (2008), who use the same model but higher spatial resolution (0.5° × 0.5°). Therefore, we find our regional model set up satisfying for simulating the mean Asian climate.

In addition to the mean climate state over Asia, the ability of our global and regional model in capturing the interannual variability of the modern Asian monsoon climate is also evaluated. More details can be found in our Paper III. Here, we would like to emphasize that modelling the interannual variation of the Asian monsoon is even more challenging than modelling its mean state. Serious biases in simulating the
interannual variation may exist, even though the mean Asian monsoon climate is satisfyingly reproduced. For instance, a bias in simulating the relation between the ISM and EASM on interannual scale has been noticed in our models. As shown in Table 3, although our models correctly describe the correlation between monsoon rainfall over Indian and northern China, they simulate a too strong correlation between ISM and EASM circulation indices compared to the reanalysis data. Because of this bias, we only analysed the interannual variation of the ISM rather than both of them in this study.

3. Summary of the original publications

3.1 Paper I

The focus of Paper I is to provide a realistic simulation of the mean Asian monsoon climate in the Late Miocene using the regional climate model to better understand its spatial pattern and driving mechanisms. Regional model simulations for the present day (i.e. CTRL) and the Tortonian (i.e. TORT) were performed and analysed. The details on their model setup can be found in Sect. 2.3.1. In this study, we found that the large-scale monsoon circulation changes between the Tortonian and present-day regional model runs (i.e. TORT-CTRL) are generally consistent with that of the driving global model runs (i.e. GTORT-GCTRL), suggesting a stronger-than-present winter monsoon but weaker-than-present summer monsoon circulation in both India and East Asia in the Late Miocene (Fig. 2 and 5 in Paper I). Precipitation changes, however, exhibit more regional differences in the regional model than in the global model experiments (Fig. 4 in Paper I). This can be attributed to the better representation of the convective activity over the monsoon region in the regional model (Fig. 8 in Paper I). The Köppen climate classification

<table>
<thead>
<tr>
<th>AIMR-NCprec</th>
<th>AIMR-SCprec</th>
<th>IMI-SWI</th>
<th>IMI-SWII</th>
<th>WY-SWI</th>
<th>WY-SWII</th>
</tr>
</thead>
<tbody>
<tr>
<td>GCTRL 0.43</td>
<td>-0.18</td>
<td>0.49</td>
<td>-0.27</td>
<td>0.21</td>
<td>-0.13</td>
</tr>
<tr>
<td>CTRL 0.42</td>
<td>-0.07</td>
<td>0.4</td>
<td>-0.03</td>
<td>0.23</td>
<td>-0.04</td>
</tr>
<tr>
<td>NCEP 0.42</td>
<td>-0.03</td>
<td>0.03</td>
<td>-0.13</td>
<td>0.01</td>
<td>-0.12</td>
</tr>
</tbody>
</table>

Fig. 3. Taylor diagram displaying a spatial statistical comparison of simulated climate over Asian monsoon area (i.e. the regional model domain) by GCTRL and CTRL with the modern observation data (CRU TS 2.1 dataset). (a) Temperature, (b) Precipitation.
based the output of TORT reveals that while the modern-like monsoon climate may have existed or even stronger than at present over southern China (most of the India), it may have not yet fully established over northern China (northwestern India) in the Tortonian due to the weak monsoon rainfall there (Fig. 7 in Paper I). The out-of-phase changes in the monsoon rainfall between northern and southern China may resolve the apparent paradox in the proxies from the two regions that indicate opposite trend of monsoon changes in the Late Miocene (Fig. 9 in Paper I).

To evaluate the relative importance of the global climate and regional tectonics for the Late Miocene monsoon evolution, two sensitivity experiments (i.e. TORTPD and PDTORT) were further performed and analysed in this study (Sect. 2.3.2). It is shown that the stronger-than-present winter monsoon in TORT can be attributed to both the stronger mid-latitude westerly flow in the global forcing and the lower northern TP that allows more westerly wind flowing through the north of TP to East Asia (Fig. 11 in Paper I). The regional difference of the monsoon rainfall changes in the Late Miocene may have particularly resulted from the combined effect of the global climate and regional tectonics (Fig. 12 in Paper I). While the lower TP and surrounding orography favours a weak summer monsoon circulation that may have led to drier conditions in northern China and northwestern India, the warmer global climate (especially higher SST) may have promoted moisture advection from the ocean, and thus the precipitation over southern China and India.

3.2. Paper II

In Paper II, we took a further step to investigate the possible effects of regional tectonics (i.e. mountain uplift) on the Asian summer monsoon changes in the Late Miocene. Previous global and regional modelling studies have mostly focused on the climatic effect of the uplift of the whole Asian mountains (i.e. bulk mountain uplift). This, however, is shown to be of less relevant to any specific geological periods, because lines of evidence have suggested more a diachronous growth of different part of the Asian orography (i.e. regional mountain uplift) than bulk mountain uplift. To explore how the regional mountain uplift may have modified the Asian climate in the Neogene, we performed a set of sensitivity experiments with successively presence of the southern TP (MsTibet), central TP (McTibet), northern TP (MnTibet), Zagros Mountains (MZagros) and Tianshan Mountains (MTian), in our regional model (see Sect. 2.3.3 and Paper II). This generally mimics the growth of the Asian orography in the Neogene (Fig. 1 in Paper II).

We found that different from the effect of bulk mountain uplift that enhances both the ISM and EASM, regional mountain uplift can lead to an asynchronous development of the ISM and EASM. The ISM is primarily intensified by the presence of the southern TP and Zagros Mountains due to the “thermal insulation” (Boos & Kuang 2010) and mechanical blocking effect, respectively (Fig. 13a and c in Paper II). In contrast, the EASM is mainly strengthened by the presence of the central TP, northern TP and the Tianshan Mountains owing to their diabatic heating effect. The elevated heating of these mountain regions can give rise to a low-level cyclonic flow around the TP. This facilitates the warm advection from the south of the TP that promotes the subtropical rain front over East Asia (Fig. 13b in Paper II), while suppresses the low-level monsoon vorticity over India, and thus the ISM.

It is argued that the asynchronous response of the ISM and EASM to regional mountain uplift may provide new insights to the monsoon evolution in the Neogene. For instance, lines of
evidence have suggested an increase (decrease) of monsoon rainfall over India (northern China) from the early Miocene to the Middle Miocene, and a decrease (increase) of monsoon rainfall over India (northern China) from the early Late Miocene to the late Late Miocene. These anti-phase changes between ISM and EASM might have been particularly associated with the uplift of the southern TP and northern TP during these periods (Fig. 12 in Paper II). In addition, the uplift of the Zagros Mountains might have been crucial for the onset of the summer monsoon over southwestern India and Pakistan (Fig. 6 in Paper II), where important monsoon proxies were preserved. The active mountain building of the Zagros Mountains around 5-3 Ma might have also contributed to the strengthening (demise) of the ISM (EASM) at ~4 Ma.

3.3. Paper III

In Paper III, we tentatively explored the interannual variability of the ISM in the Late Miocene by analysing the 110-year simulations of GTORT and TORT. It is found that the standard deviations of different ISM monsoon indices in our Tortonian runs are all similar to or even stronger than in the present-day control runs (Table 2 in Paper III). This indicates a similar or even stronger interannual variability of the large-scale ISM in the Late Miocene. The spectral analysis of the ISM index (IMI) further reveals that the variation of the ISM may have had a shorter dominant cycles (2.6-2.7 year/cycle) in the Late Miocene compared to today (Fig. 3 in Paper III). The spatial pattern of the summer rainfall and wind changes associated with the ISM variation in the Late Miocene is shown to have been analogous to today, except that the rainfall variation over northern China and northwestern India might have been weaker than present due to the weak mean summer monsoon in these regions (Fig. 4 and 5 in Paper III). Compared to the global model results, our regional model provides little adding information on the spatial structure of the interannual variations of the summer rainfall and wind (cf. Fig. 4 and 5 in Paper III). This implies a dominant control of the external global forcing on the interannual variation of the Asian summer monsoon, and the internal processes within the monsoon region plays a minor role.

It is suggested that the strong ISM interannual variability may have been maintained by the stronger-than-present El Niño-Southern Oscillation (ENSO) in the Late Miocene (Table 4 and Fig. 7 in Paper III). Meanwhile, the extratropical influence, such as the summer North Atlantic Oscillation (NAO) and the winter/spring snow cover, on the ISM may have been weaker (Table 6 and Fig. 10 in Paper III). This allows a stronger modulation of ENSO on ISM that may have resulted in an interannual variation of ISM at a period closer to that of ENSO in the Late Miocene. We note that the strong ENSO in the Late Miocene might have been caused by the combined effect of the open Panama and Indonesian Seaways, the weakened thermohaline circulation and the lower global orography at that time. Various state-of-the-art CGCMs have demonstrated a strong ENSO variability in other warm periods of the deep past and also in the future. This indicates that strong ENSO and ENSO-ISM teleconnection might be a robust pattern in the global warming climate. As a result, a strong interannual variability of the ISM might have appeared in the greenhouse world of the deep past, even though the mean ISM might have not yet fully developed compared to the present day.

4. Discussion

4.1. Spatial pattern of the Asian monsoon climate in the Late Miocene
4.1.1. Distribution of monsoonal climate

A striking spatial pattern of modern monsoon climate is the spread of monsoon (humid) climate over subtropical East China that is distinct from the non-monsoon (arid) climate in its western counterpart (see Fig. 7a in Paper I). The transition from the humid to arid climate is usually depicted by a southwest-northeast oriented line over the Loess Plateau (referred to as “monsoon pattern”), which is very sensitive to the relative strength of the East Asian summer and winter monsoon.

It has been suggested that, when the East Asian monsoon climate was weak or not yet established, the arid climate might have covered the whole northern China due to the dominant influence of the subtropical subsidence at planetary scale (Guo et al 2008, Sun & Wang 2005). The boundary between humid and arid climate, therefore, should have been more zonally directed (referred to as “planetary pattern”). When such planetary pattern transformed into the monsoon pattern in the geological past remains controversial. Using palaeobotanic and lithological data from China, Sun and Wang (2005) and Guo et al (2008) demonstrated that the monsoon pattern may have appeared in the Latest Oligocene and Early Miocene. However, recent climate reconstruction based on mammal and plant fossil data revealed that the planetary pattern might have persisted in the Middle Miocene and early Late Miocene before being replaced by the monsoon pattern at 8-7 Ma due to the strengthening of the EASM (L.P.Liu et al 2009, Y.S.Liu et al 2011).

Climate models have been used to explore the evolution of the monsoon pattern in these periods (Zhang et al 2007a,b). To delineate the extent of the monsoon climate in the climate model, various criteria have been applied. They are summarized in Table 4. In general, all these methods are according to the basic feature of the monsoonal climate, i.e. the large seasonality in either wind direction and strength (Li & Zheng 2003, Zhang et al 2007a) or in precipitation (Huber & Goldner 2012, Wang & Ding 2006, Wang & LinHo 2002). Although these methods have been validated in modern climate, several caveats should be considered before applying them to palaeoclimate modeling results. (1) The model biases in simulating wind and precipitation may affect the results. (2) Methods using threshold value of precipitation to distinguish the monsoonal climate is based on present-day climate conditions. The threshold values in the deep past might have been quite different from the present.

Table 4. Summary of different methods used for defining the monsoon climate region.

<table>
<thead>
<tr>
<th>References</th>
<th>Methods</th>
</tr>
</thead>
<tbody>
<tr>
<td>Li and Zheng (2003)</td>
<td>Relative difference between winter and summer wind vector (i.e. SNS) larger than 0.2, $S_{\text{NSF}} = F_1 - F_7$ (Li &amp; Zheng 2003)</td>
</tr>
<tr>
<td>Zhang et al (2007a)</td>
<td>Precipitation larger than 1.5 mm/day. (Zhang et al 2007a)</td>
</tr>
<tr>
<td>Zhang et al (2007b)</td>
<td>The difference between climatological pentad mean and January mean precipitation exceeds 5 mm/day (or local January mean rainfall rate). (Zhang et al 2007b)</td>
</tr>
<tr>
<td>Wang and LinHo (2002)</td>
<td>Annual range (i.e. local mean summer (MJIAS) minus winter precipitation) exceeds 2 mm/day; Local summer precipitation exceeding 55% of the annual total. (Wang and LinHo 2002)</td>
</tr>
<tr>
<td>Wang and Ding (2006)</td>
<td>Ratio of summer (MJIAS) to annual precipitation larger than 70%. (Wang and Ding 2006)</td>
</tr>
<tr>
<td>Huber and Goldner (2012)</td>
<td>Köppen climate classification: (1) Arid climate: $P_{\text{ann}} &lt; 10P_{\text{th}}$; (2) Winter-dry climate: $P_{\text{ann}} &lt; P_{\text{min}}$ and $P_{\text{ann}} &gt; 10P_{\text{min}}$ (Kottek et al. 2006)</td>
</tr>
<tr>
<td>This study and Z.S. Zhang et al (2012)</td>
<td>Köppen climate classification: (1) Arid climate: $P_{\text{ann}} &lt; 10P_{\text{th}}$; (2) Winter-dry climate: $P_{\text{ann}} &lt; P_{\text{min}}$ and $P_{\text{ann}} &gt; 10P_{\text{min}}$ (Kottek et al. 2006)</td>
</tr>
</tbody>
</table>
day due to the changes in other climate conditions, such as temperature. (3) Methods using seasonality of wind direction and strength may fail to distinguish monsoon climate in some regions. For instance, some arid areas in Central Asia exhibiting significant seasonal wind changes may be mistakenly divided into monsoonal climate (Zhang et al 2007a). In our study, we chose the Köppen climate classification, an approach based on both temperature and precipitation, to depict the extent of the monsoon (humid) climate in our Late Miocene model experiments. It is found that the boundaries between the winter-dry (Dwb) and the arid-steppe (BSk) climate in this classification can well capture the present-day transition from monsoon to non-monsoon climate in northern China (Fig. 7a in Paper I). An advantage of this method is that temperature is incorporated in defining the climate types (e.g. BSk and Dwb) (see Table 4). To avoid the influence of our model biases (see Sect. 2.5) on the results, we used the modern observation as baseline and added the anomalies of temperature and precipitation between our Tortonian run (i.e. TORT) and the present-day control run (i.e. CTRL) to the baseline dataset to generate climate classification for the Late Miocene.

Our results show that northern China is mostly covered by arid-steppe (BSk) in our Tortonian run (Fig. 7b in Paper I). This is consistent with the studies by L.P.Liu et al (2009) and Y.S.Liu et al (2011) supporting that the modern-like monsoon pattern may have not established in northern China in the Tortonian. Further sensitivity experiments indicate that the uplift of the northern TP during the Late Miocene and Pliocene might have been essential for the formation of monsoon pattern in northern China (Fig. 4d). In addition, the small-scale orography in northern China, such as the Taihang Mountains, might have also played a role in shaping the modern monsoon pattern (cf. Fig. 4f and g). Our results is in contrast with previous model studies by Zhang et al (2007a, 2007b), which imply much earlier establishment of the monsoon pattern in northern China in response to tectonic changes. This can be due to the model biases and different methods used for delineating the monsoon climate in their studies (Table 4).

Compared to the distribution of monsoon climate in northern China, the monsoon climate in India (as denoted by winter-dry climates: Aw and Cwa in Köppen classification) might have reached its modern extent in the Tortonian (Fig. 7b in Paper I). However, the foreland basin over northern Pakistan is covered by summer-dry climate (Csa) in TORT (Fig. 7b in Paper I), suggesting the monsoon climate in this region may have initiated later than the Tortonian. This is in accordance with the monsoon proxies from this area (Quade et al 1989, Sanyal et al 2010). It is shown that the extent of monsoon climate in India can also change significantly in response to mountain uplift (Fig. 4). It may retreat to southeastern India when no mountains are present (Fig. 4a), while it may achieve its modern extent with the presence of the southern TP (Fig. 4b), or both the TP and the Zagros Mountains (Fig. 4e). This explains the present-day extent of the monsoon climate over India in our Tortonian run (TORT).

4.1.2. Development of arid climate
It is commonly thought that the arid climate in central Eurasia may have co-evolved with the Asian monsoon climate in the Neogene. Two opinions have prevailed about their relationship in these periods. Most modeling studies advocates a synchronous strengthening in the intensity of the monsoon and arid climates, primarily controlled by the regional tectonics, such as the uplift of the TP (An et al 2001, Broccoli & Manabe 1992) and the retreat of Paratethys (Ramstein et al 1997, Zhang et al 2007a). Most proxy studies,
however, reveal an enhanced (weakened) arid (monsoon) climate, closely associated with the global cooling (Lu et al 2010, Miao et al 2011). Recent review on the available proxies by Miao et al (2012) indicates regional differences in the development of arid climate throughout Middle and Late Miocene, pointing to a combined effect of global cooling and regional mountain uplift in producing these regional differences.

The spatial difference in the development of arid climate has also been observed in our Paper I. In particular, we emphasize the differences between subtropical (south of 40 °N) and mid-latitude desert (north of 40 °N). It is shown that the large extent of desert (mostly the subtropical desert) may have existed in western Asia, northern TP and Tarim basin in the Tortonian (Fig. 7b in Paper I). In contrast, the desert climate over Central Asia (east of the Tianshan Mountains) and over Mongolia (mainly the mid-latitude desert) may have been less extensive in the Tortonian than at present (Fig. 7a in Paper I). The existence of

arid climate over northern TP and its north is consistent with the δ¹⁸O records from Linxia Basin (Dettman et al 2003) and the carbon isotope signature in the aeolian deposition from central North Pacific (Jia et al 2012). The earliest dust deposition in the Tarim Basin has been dated at 8-7 Ma (Sun et al 2009, Zheng et al 2010). This also implies the initiation of desert in this region no later than 8 Ma. Our results, however, contradict Ma et al (1998, 2005) who claim humid conditions in these regions in the Tortonian. The relatively humid conditions of Central Asia (west of the Tianshan Mountains) agrees with a pollen study by Sun and Zhang (2008) and the low hypsodonty of this period (Fig. 10a in Paper I).

It is noticed that the regional mountain uplift might have been particularly responsible for the regional difference in the development of the arid climate in central Eurasia. As illustrated in Fig. 4, most of the Asian continent between 20-40 °N, including half of India, is covered by desert climate (mostly the subtropical desert), when all
the Asian orography is removed (Fig. 4a). This can be attributed to the dominant influence of the planetary-scale subtropical subsidence in this region and its far distance from the ocean. The presence of southern TP greatly enhances the ISM, resulting in the disappearance of the arid climate over India. It also enhances the aridity north of the southern TP by blocking moisture transport from the Indian ocean, but has little influence on the extent of desert area in this region (Fig. 4b). The presence of the central and northern TP leads to the advancement (retreat) of monsoon (arid) climate over northern China, meanwhile facilitates the expansion of desert climate more to the mid-latitude Mongolia (Fig. 4c and d). This can be explained by the stationary waves induced by the mountain uplift that inhibiting precipitation in winter and spring, and the intensified subsidence in summer in responsible to the strengthened convergence over northern TP (Broccoli & Manabe 1992, Sato 2005). In contrast, the presence of the Zagros Mountains mainly contributes to the appearance of the mid-latitude desert over Central Asia (west of the Tianshan Mountains) (Fig. 4e). This can be ascribed to the monsoon-desert mechanisms (Rodwell & Hoskins 1996, Wu et al 2009), which states that the enhanced monsoon convection over India can cause an anticyclonic flow that favours descent motion over Central Asia by Rossby-wave response. In addition, the mechanical effect of the Zagros Mountains may also play a role in maintaining the anticyclonic flow over Central Asia (Bollasina & Nigam 2011). The presence of the Tianshan Mountains allows the spread of humid climate in the centre of desert climate along the mountain ranges because of the orographic rainfall effect (Fig. 4f), but has small influence on the large-scale extent of the desert climate.

We note that although global cooling has been widely invoked to explain the overall drying trend of the Asia interior in the Neogene, the mechanisms of such effect have not been well demonstrated by climate models (Liu et al 2013). In particular, how the global cooling modifies the water balance, especially the temperature related evaporation, in the arid area, has been overlooked. A recent modeling study by Shi et al (2011) has suggested that the global cooling might have minor influence on the extent of dust source region (i.e. desert climate) since mid-Pliocene. Instead, the uplift of the TP is found to be more fundamental. It is also noticed in our results that the warmer climate in the Tortonian might have not produced a uniformly wetter conditions in central Eurasia. Pronounced increase of precipitation is only observed in the mid-latitude region due to the strengthened storm track (Fig. 12 in Paper I). Further modelling studies are required to better evaluate the influence of global climate on the arid climate changes in this period.

4.2. Summer monsoon in the Late Miocene

Summer monsoon is the most prominent component of the modern Asian monsoon system. Most of the monsoon proxies from the Late Miocene recorded rainfall intensity or seasonality, which is primarily controlled by summer monsoon strength. However, no consensus has been reached on the relative strength of the summer monsoon in the Late Miocene. Some proxies suggest weaker-than-present summer monsoon in the Tortonian and intensification later at 8 Ma due to the uplift of the TP (Chen et al 2003, Forstelius et al 2002, Huang et al 2007). Some studies indicate a stronger-than or similar-to present summer monsoon strength in the Tortonian owing to the warmer global climate (Dettman et al 2001, Hoorn et al 2000, Jiang & Ding 2009, Steinke et al 2010). We found that several features of the summer monsoon manifested in our Tortonian model runs might provide useful clues.
to resolve these apparent paradoxes in the summer monsoon proxies from the Late Miocene.

4.2.1. Decoupling of summer monsoon circulation and precipitation

It is indicated in our global Tortonian run (GTORT) that while the summer monsoon circulation over India might have been weaker in the Tortonian, monsoon rainfall over the whole Indian subcontinent might have been stronger than today, due to the higher moisture loading in the atmosphere maintained by the high SST over the Indian Ocean (see Fig. 2b and d in Paper I). Although our regional model run (TORT) does not support wetter conditions over the whole Indian Subcontinent as in GTORT, there are still some regions, such as the western coast and the southern tip of India, showing higher rainfall in TORT. This suggests a decoupling of the large-scale monsoon circulation and monsoon rainfall at least in some regions over India in the Tortonian. Such decoupling might offer an explanation to the discrepancy between the proxy from the Arabian Sea that suggests a weak ISM circulation (Huang et al 2007) and the proxies from northern India that claims a more humid conditions in the Tortonian (Dettman et al 2001, Hoorn et al 2000). Note that the decoupling between the ISM circulation and precipitation has also been found in multiple CGCM simulations for the future (Kripalani et al 2007b, Paeth et al 2008). This points to the importance of distinguishing circulation and rainfall when interpreting the ISM strength in the Late Miocene as well as other periods.

4.2.2. Regional difference in monsoon rainfall

It is suggested in TORT that the monsoon rainfall changes might have exhibited pronounced regional difference in the Tortonian (Fig. 4 in Paper I). In particular, precipitation over southern China (southern India) might have been stronger-than-present in the Tortonian, which is in contrast to the weaker-than-present monsoon rainfall over northern China (northwest India). Such regional differences are inherent features of the Asian summer monsoon systems, which have been widely observed in modern monsoon changes on interannual scale (Ding 1992, Gadgil 2003). The anti-phase changes of monsoon rainfall between northern and southern China can be linked to the displacement of the western North Pacific subtropical high and the associated rain front along its northwestern flank which is the primary source for the EASM rainfall (Chang et al 2000, Liu et al 2008, Wang et al 2013). In contrast, the inverse-phase changes of monsoon rainfall between northwestern and southern India can be attributed to the shift in the mean position of the intertropical convergence zone (ITCZ) (Fluteau et al 1999, Gadgil 2003). The southward shift of the mean position of the western North Pacific subtropical high (ITCZ) explains the regional differences of monsoon rainfall changes between northern China (northwestern India) and southern China (southern India) in the Tortonian (Fig. 6d and Fig. 8b in Paper I).

The proxies retrieved from southern China have revealed a stronger-than-present monsoon rainfall in the Tortonian (Clift et al 2008, Steinke et al 2010). This is distinct from the monsoon proxies from northern China, which suggest a weaker monsoon rainfall in this region (An et al 2001, Fortelius et al 2002). Earlier studies have used both results as indicators for the EASM strength, therefore arrived at contradictory conclusions for the EASM strength in the Tortonian (Clift et al 2008, Steinke et al 2010). However, according to our modelling results, such inverse-phase variation of the monsoon rainfall between northern and southern China may actually point to a coherent change in EASM strength, i.e. a weaker EASM in the Tortonian (based on our definition of the EASM strength).
4.2.3. Asynchronous development of the ISM and EASM

Many palaeomonsoon studies have naturally held the idea that the ISM and EASM developed synchronously throughout the Miocene and Pliocene (An et al 2001, Clift et al 2008). This is understandable given that the thermal contrast between Eurasian continent and surrounding oceans is the primary forcing for both the ISM and EASM. Climate models have also demonstrated that both the EASM and ISM can be intensified by the uplift of the TP (Kutzbach et al 1993, Liu & Yin 2002, Song et al 2010a), further supporting the synchronous development of both monsoons in these periods. However, there are also evidence suggesting an asynchronous development of the ISM and EASM in the Miocene and Pliocene. For instance, it has been suggested that the EASM might have been initiated in Early Miocene (Guo et al 2008, Sun & Wang 2005), while the earliest evidence for the onset of the ISM might be traced back to Eocene (Huber & Goldner 2012). The mechanisms for such asynchronous development of the summer monsoons have been less understood.

The asynchronous variation of the ISM and EASM has not been observed in modern monsoon changes on interannual scale (Wang et al 2001a). Instead, a close correlation between summer rainfall in northern China and India has been found (Ding & Wang 2005) (see also Table 3), suggesting a more synchronous change of the ISM and EASM. This is attributed to the teleconnection between the upper level positive geopotential anomaly over west-Central Asian induced by the ISM convection and that over northeastern Asia through wave-train response in summer (Ding & Wang 2005). The barotropic positive geopotential anomaly over northeastern Asia promotes low-level southeasterly wind to its south and southwestern side (i.e. northern China), therefore enhancing precipitation there. In contrast, the regions located to the central and eastern part of the positive geopotential anomaly (i.e. southern Japan) is dominated by subsidence, thus reduced rainfall.

There is several lines of evidence that would support asynchronous changes of the summer rainfall over India and northern China. For instance, Zhang (2001) found an inverse relationship between the water vapour transport from the Bay of Bengal to India and to East Asia on interannual scale. Chen et al (2011) reported an asynchronous change of summer rainfall over India and northern China in response to obliquity forcing on orbital scale. Interestingly, they also link this to the teleconnection between the upper-level geopotential anomalies over India and East Asia, which is similar to that proposed by Ding and Wang (2005). The difference is that the positive geopotential anomaly over northeastern Asia in Ding and Wang (2005) is more centred over northern China in Chen et al (2011) which suppresses (rather than promotes) the summer rainfall in northern China. Such discrepancy in these two studies may result from the biases of the model used by Chen et al (2011) in representing the observed teleconnection pattern as shown by Ding and Wang (2005). It may also implies that the teleconnection pattern between India and northern China may have altered in the geological past due to different physical boundary conditions, therefore leading to different relationship between the summer rainfall over Indian and northern China.

On tectonic scales, it has been noticed that the changes of vegetation over the Sahara from grassland to desert in the Late Miocene may have induced a decrease of ISM rainfall but an increase of EASM rainfall (Micheels et al 2009). It has also been argued in recent studies that the ISM and EASM may have disparate responses to the mountain uplift in Asia (Boos & Kuang 2010,
R. Zhang et al. (2012). In our paper II, we further demonstrate this possibility and propose that the regional mountain uplift may have induced asynchronous changes of the ISM and EASM in the Miocene and Pliocene. For instance, the uplift of the southern TP may have been responsible for the strengthening (weakening) of the ISM (EASM) from Early to Middle Miocene (Fig. 12 in Paper II). The weakening (strengthening) of the ISM (EASM) from the early Late Miocene (Tortonian) to the late Late Miocene might have been associated with the uplift of the northern TP. In addition, the significant uplift of the Zagros Mountains around 5-3 Ma may have contributed to the strengthening (weakening) of the ISM (EASM) at 4 Ma.

Similar to Zhang (2001), we also found an inverse-phase change between the water vapour transport from the Bay of Bengal to India and to East Asia in response to regional mountain uplift (figure not shown), which contributes to the asynchronous development of the ISM and EASM. However, limited by our regional model domain, we do not see the upper level teleconnection pattern between India and northern China existing in our experiments. Further studies with wider regional model domain would be able to detect the contribution of such teleconnection on the asynchronous response of the ISM and EASM to the regional mountain uplift.

4.3. Strong winter monsoon in the Late Miocene

Compared to the summer monsoon, winter monsoon in the Late Miocene has been less extensively studied and hence more uncertain. Some studies suggest a weaker than present EAWM in the Late Miocene and an intensification of the winter monsoon at 8-7 Ma or later (An 2000, Jacques et al. 2013, Jia et al. 2003, Jiang & Ding 2010, Rea et al. 1998, Wan et al. 2007, Wang et al. 2003). This has been attributed to either the lower TP or the warmer global climate (particularly in the high latitudes) in the Late Miocene. Both mechanisms can result in a weaker Siberian High and thus winter monsoon wind over East Asia. Most of the winter monsoon studies use aeolian deposition rate to infer the winter monsoon strength for the Late Miocene (Jia et al. 2003, Jiang & Ding 2010, Rea et al. 1998, Wan et al. 2007). This proxy, however, does not exclusively modulated by winter monsoon wind, but also influenced by the aridity and wind speed over the dust source region (e.g. Central Asia), and the vegetation and precipitation over the dust.
capture region (e.g. northern China and North Pacific) (Shi et al 2011, Yue et al 2009). In addition, dust storms occur more frequently in spring than in winter (Molnar et al 2010), indicating dust transportation and accumulation may not be directly related to winter monsoon. These factors diminish the reliability of aeolian deposition rate in reflecting the winter monsoon strength.

In comparison to dust accumulation rate, we propose that winter surface temperature might be a more direct and hence robust indicator for the winter monsoon strength. The outbreak of winter monsoon wind from the cold north (e.g. Siberian High) is always associated with a drop of temperature over East Asia due to strong cold advection. Using reconstructed (winter) temperature from fossil snails as an indicator for the winter monsoon strength, Li et al (2008) found a strong winter monsoon and winter monsoon dominated climate over western Loess Plateau before 6 Ma. Recent quantitative climate reconstruction based on plant fossils also revealed a colder winter temperature over southern China in the Later Miocene or even earlier periods (Jacques et al 2011, Quan et al 2012, Xing et al 2012, Q.Q. Zhang et al 2012), suggesting a strong winter monsoon in this region. This is in contrast to previous studies which implied a relatively weak winter monsoon in the Late Miocene based on aeolian deposition rate.

Both our global and regional model experiments support a stronger winter monsoon over East Asia and India in the Tortonian than at present (Fig. 2 and 5 in Paper I). Similar results have also been found in recent mid-Pliocene climate simulations using physical boundary conditions analogous to our Tortonian runs, except that the strong winter monsoon wind is only observed in southern China and India, while the winter monsoon wind over northern China is weakened (Zhang et al 2013). We found that the regional tectonic changes may have been particularly responsible for the strong winter monsoon in the Late Miocene (Fig. 11 in Paper I). Especially, the lower northern (southeastern) TP allows more low-level westerly (easterly) wind flowing through the region, and thus facilitates the northwesterly (northeasterly) to East China (India). This accounts for the decline of winter temperature in both East China and India (Fig. 5d).

It is interesting to note that the causes for the strengthening of the EAWM due to the lower northern TP is distinct from that for the intensification of the EAWM with the uplift of the entire TP as demonstrated in earlier studies (An et al 2001, Kutzbach et al 1993, Liu & Yin 2002). While the former is mainly driven by the strengthened westerly flow, the latter is primarily caused by the enhanced Siberian High. The dominant influence of the westerly flow in maintaining the winter monsoon in the Tortonian when
the northern TP was low is consistent with studies by Ding et al (1999) and Sun (2004), which emphasize the importance of westerly flow for dust transport in the Late Miocene. The uplift of the northern TP may have given rise to a shift from the westerly driven winter monsoon to the Siberian High driven winter monsoon since the Late Miocene.

It is also worth noting that the decrease of winter temperature over East China and India due to strong winter monsoon might have been counteracted by the warmer global climate in the Late Miocene (Fig. 5c). This explains the generally warmer winter temperature over the Asian continent in our global Tortonian run (GTORT) (Fig. 5a) and the less decrease of winter temperature over India and East China in our regional Tortonian run (TORT) than the pure effect of the regional tectonic changes (cf. Fig. 5b and d). Since the global warming effect is stronger in northern China, the strong winter monsoon in the Late Miocene may not be reflected by the winter temperature changes over north China. Instead, it may be better detected by the winter temperature changes over southern China and India, where the global warming effect is less pronounced. It is also found that our regional model simulates a stronger decrease of winter temperature over India and southern China in the Tortonian than the global model (cf. Fig. 5a and b). This is in better agreement with the temperature reconstructions from proxy data (see Sect. 4.6), therefore suggesting an improved representation of the winter monsoon in the Tortonian by our regional model experiment.

4.4. Strong Interannual variation of the ISM in the Late Miocene

Little has been known about the interannual variation of the summer monsoons in the Late Miocene and its association with the mean state evolution of the summer monsoons, due to the lack of annually resolved summer monsoon records from this period. The interannual variation of the ISM is closely associated with ENSO in the tropical Pacific (Gadgil 2003, Ju & Slingo 1995, Wang et al 2001a, Webster et al 1998). The lighter (heavier) than normal ISM rainfall normally corresponds to positive (negative) phase of ENSO, i.e. the El Niño (La Niña) conditions. Some proxy studies have claimed a permanent El Niño-like condition in the tropical Pacific before the mid-Pliocene (Fedorov et al 2010, Molnar & Cane 2002, Wara et al 2005). If such permanent El Niño-like condition took place, a weak interannual variability of the ISM is expected in the Late Miocene. However, recent geological evidence revealed a persistent interannual ENSO variability in the Pliocene and the preceding periods (Galeotti et al 2010, Ivany et al 2011, Scroxton et al 2011, Watanabe et al 2011). If such modern-like ENSO variability existed in the Late Miocene, the interannual variability of the ISM might have resembled its present-day state.

It is found in our global Tortonian run that the ISM interannual variability might have been strong or even stronger than at present in the Late Miocene. Moreover, there was a trend towards shorter and more frequent oscillation (about 2-3 year/cycle) of the ISM strength in the Late Miocene compared to today (5-8 year/cycle) (Fig. 3 in Paper III), indicating a strong quasi-biennial oscillation in the Late Miocene. This result might be testable by proxy data in the future. We also observe a strong ENSO and ENSO-ISM teleconnection in our Tortonian global run (Fig. 7 and 8 in Paper III), which supports a persistent ENSO variability in the Late Miocene (Galeotti et al 2010, Watanabe et al 2011). We propose that the strong ENSO and ENSO-ISM teleconnection might have resulted in a stronger and shorter ISM variability in the Late Miocene. In addition, the extratropical forcing, such as the teleconnection
between summer NAO and the ISM, might have been weak in the Late Miocene (Table 6 in Paper III). This could further facilitate the interaction between ENSO and ISM, therefore favour a biennial oscillation of the ISM (Meehl 1994). We notice that several factors might have contributed to the strong ENSO in the Late Miocene, such as the open Panama and Indonesian Seaway (von der Heydt & Dijkstra 2011), the weak thermohaline circulation (Lu et al 2008), the lower global orography (Kitoh 2007) as well as the warming global climate in general (Fu & Lu 2010). This indicates that a strong ENSO might be a robust pattern in the Late Miocene climate.

Compared to the changes in its temporal structure, the spatial pattern of the ISM interannual variation in the Late Miocene might have remained similar to today according to our modelling results (Fig. 4 in Paper III). However, some regional scale differences might have occurred owing to the changes in the mean summer monsoon state, such as the southward shift of the zonal elongated maximum rainfall variability over East China and the weakening of the rainfall variability over northwestern India (Fig. 5 in Paper III). These differences are better depicted by our regional model and may also be tested by proxy data in the future.

It is worth mentioning that the strength of the ISM interannual variability might have been decoupled with its mean state in the Late Miocene (Table 2 in Paper III). Such decoupling is related to the fact that the interannual variability of the ISM is more controlled by external forcings, e.g. ENSO, rather than internal feedbacks within the monsoon region. This is different from other climate system, such as NAO, which interannual variability is closely linked to its mean state. The positive mode of the mean state NAO has been found to exhibit weaker NAO variability on interannual scale (Paeth et al 1999), which is also observed in our results (Table 5 in Paper III). This suggests a strong control of the internal feedbacks related to the mean state NAO in determining its interannual variability. The strong interannual variability of the ISM in our Tortonian experiments exhibits great resemblance to that in the future climate projections (Table 7 in Paper III). This implies that the Late Miocene might be a good analogue of the future climate in terms of the Asian monsoon variability. Finding more proxies indicating the monsoon variability in this period will improve our understanding of the Asian monsoon evolution not only in the Late Miocene but also in the future.

4.5. Global and regional forcings for the Late Miocene monsoon changes

Both the global (e.g. global cooling) and regional (e.g. uplift of the TP) forcings have been invoked to explain the monsoon evolution in the Miocene and Pliocene (An 2000, Clift et al 2008, Ge et al 2012, Miao et al 2012). But the relative importance of these two forcings and their interaction in shaping the monsoon climate in the Late Miocene remain controversial. Regional climate model provides an unique tool to investigate this problem. The effect of regional forcings (e.g. regional mountain uplift) can be explicitly demonstrated if only the physical boundary conditions in the regional model are modified and the lateral boundary conditions are kept the same. On the other hand, the effect of the global forcings can be generally inferred if the lateral boundary conditions are altered while the physical boundary conditions in the regional model remain unchanged. Using this technique, we have examined the influence of global and regional forcings on different aspects of the monsoon climate in the Late Miocene. The results are summarized in Table 5.

4.5.1. Effects of global forcings
It has been proposed in previous studies that warmer conditions in the Miocene and Pliocene might have maintained a strong summer monsoon and thus humid conditions over the monsoon regions (Clift et al 2008, Passey et al 2009). This can be related to two possible mechanisms: (1) the warming over northern Eurasia that enhances the land-sea thermal contrast in summer and thus the summer monsoon; (2) the higher SST over ocean that promotes moisture supply for the monsoon rainfall. We found that the second mechanism might work in our model and is responsible for the wetter conditions over Indian and southern China in our Tortonian runs (TORT and GTORT) (Fig. 2 and 4 in Paper I). But we see no evidence that the first mechanism would have worked in the Later Miocene. In particular, the pronounced warming over northern Eurasia in our Tortonian runs does not give rise to a strong summer monsoon wind over northern China that would allow the penetration of monsoon rainfall into this region (Fig. 4 in Paper I).

Besides warmer global conditions, the changes in the strength of mid-latitude westerlies and storm track in global climate may have also exerted profound impact on the monsoon climate in the Late Miocene. The stronger westerly flow and storm track observed in our Tortonian global run contribute to a stronger winter monsoon in northern China and wetter conditions over mid-latitude central Eurasia. There is also evidence showing other factors in the global climate system might have affected the Asian monsoon in the Late Miocene through distinct processes, such as the vegetation over Sahara (Micheels et al 2009) and the open Panama Seaway (Motoi & Chan 2010). These effects are mixed with each other in our global Tortonian run, and therefore difficult to separate in this study.

4.5.2. Effects of regional forcings

The strong impact of regional tectonics, e.g. uplift of the TP, on the Asian monsoon development have been well demonstrated by climate models (Hahn & Manabe 1975, Kutzbach et al 1993, Liu & Yin 2002, Prell & Kutzbach 1992) and widely used to interpret monsoon evolution in the Miocene and Pliocene (Harris 2006, Mol-
However, our understanding on the role of regional tectonics in shaping the monsoon climate in the Late Miocene is still ambiguous, due to the uncertainty of our knowledge on the tectonic history and palaeoelevation of Asian orography and also due to our incomplete grasp of the potential effect the Asian orography could exert on the monsoon climate.

The orography prescribed in our regional Tortonian run follows the prevailing evidence for the evolution of Asian orography in the Late Miocene. It points to a similar-to-present height of the southern TP but much lower northern TP (Fig. 1 in Paper II). Our results highlight the role of northern TP in determining the summer monsoon strength over northern China and the mid-latitude aridity. The lower northern TP in the Late Miocene might have suppressed the summer monsoon in northern China and reduced the aridity over Mongolia (Fig. 12 in Paper I). Such effect is consistent with other modelling studies (Boos & Kuang 2010, R.Zhang et al 2012). Interestingly, the northern TP is also an important player for the winter monsoon. As discussed in Sect. 4.3, the lower northern TP in the Late Miocene might have strengthened the winter monsoon wind over East China and India that caused the decline of winter temperature in these regions.

Several mechanisms have been advanced to explain the effect of the Asian orography on the Asian monsoon climate, especially the summer monsoon (Kutzbach et al 1993). Early studies suggest that the mountain regions act as an elevated heating source in summer that induces an thermal high pressure anomaly in the upper troposphere (Duan & Wu 2005, Duan et al 2008, Kutzbach et al 1993, Wu et al 2012). This maintains an upper level divergent flow which drives a low-level convergent flow around the TP, therefore both the ISM and EASM (referred to as “diabatic heating effect”). Recent studies, however, emphasize that the prevention of cold air in the extratropics from advecting to India by the narrow mountains chains, such as the Himalayas and the Zagros Mountains, may be essential for the ISM (i.e. thermal insulation effect) (Boos and Zhang 2010, 2013). In addition, the Asian orography as a physical barrier that forces the westerly wind overflow or circumvent the TP has also been found responsible for the monsoon rainfall over East Asia, especially in early summer (i.e. mechanical effect) (Park et al 2012). There have been controversies over the relative role of these mechanisms in affecting the summer monsoon response to mountain uplift. We found that a combination of the above mentioned mechanisms is responsible for the synchronous strengthening of the ISM and EASM with the bulk mountain uplift (Paper II). However, with regard to the regional mountain uplift, single mechanism may dominate in influencing the summer monsoon changes. For instance, the southern TP (the Zagros Mountains) exhibits strong thermal insulation (mechanical blocking) effect that mainly intensifies the ISM (Fig. 13a,c in Paper II), while the central and northern TP have strong diabatic heating effect that primarily stimulates the EASM (Fig. 13b in Paper II). As a result, we observe an asynchronous development of the ISM and EASM with regional mountain uplift.

In addition to regional tectonics, the effect of regional vegetation on the Asian monsoon has also been examined. Its influence is small in our regional model, indicating that the regional vegetation changes might have played a secondary role in the monsoon evolution in the Late Miocene. The retreat of the Paratethys has also been shown to have great impact on the monsoon development in the Miocene (Fluteau et al 1999, Zhang et al 2007b). However, its changes since the Late Miocene were relatively small (Popov et al 2006) and its effect on the Asian climate is
less uncertain. Therefore, we do not investigate its influence in more details in our study.

4.5.3 Synergy effect of global and regional forcings

It is found that the global and regional forcings for the monsoon changes in the Late Miocene may either reinforce or counteract with each other. For instance, both the global and regional forcings might have contributed to the wetter conditions over mid-latitude Central Asia and the strong winter monsoon wind over northern China (Table 5). In contrast, the regional difference in summer rainfall and winter temperature changes between northern and southern China (or India) may have particularly resulted from the competing effect of the global and regional forcings.

The combined effect of the global and regional forcings in our regional model is largely linear compared to their separate effects (Table 6). However, small synergy effect of these two forcings is detected in various monsoon indices (Table 6). The negative synergy effect of the winter monsoon index (MO) and summer monsoon indices (EIMR, WY, SWI, NCprec and SCprec) indicates a reinforced influence of the lower orography on these indices under the Tortonian global forcing compared to that under the present-day global forcing. This implies a generally higher sensitivity of the monsoon climate to the lower orography under the Tortonian global climate conditions than that under the present-day global climate conditions. Differently, the negative synergy effect of the EAWM index (WJ) indicates a weakened effect of the lower northern TP on the strengthening of the EAWM wind under the Tortonian global forcing. The relatively large negative synergy effect of the EASM index (SWII) denotes opposite effect of the lower northern TP on the monsoon convergence over central China (i.e. the lower reach of the Yangtze River valley) under the Tortonian and present global forcing. This can be attributed to the difference in the mean position of the subtropical rain front over East Asia between the Tortonian and present global forcing. In the present-day (Tortonian) global forcing, the mean position of the subtropical rain front is over northern (central) China. The lower northern TP displaces the subtropical rain front southward, therefore enhances (weakens) the monsoon convergence over central China.

4.6. Model-proxy comparison for the Late Miocene

To validate the performance of our model experiments for the Late Miocene, both the qualitative and quantitative climate reconstructions from proxy records have been used. The comparison of our modelling results with the qualitative monsoon proxies from the Late Miocene has been analysed in Paper I. In this section, we will briefly show a comparison with quantitative temperature and precipitation reconstruc-

Table 6. Effect of Tortonian global (TORTPD-CTRL) and regional forcings (PDTORT-CTRL) on the monsoon changes and their synergy effect. NCprec and SCprec denote summer precipitation over northern China (105-120 °E, 34-42 °N) and southern China (105-120 °E, 22-30 °N), respectively.

<table>
<thead>
<tr>
<th>MO</th>
<th>WJ</th>
<th>EIMR</th>
<th>WY</th>
<th>SWI</th>
<th>SWII</th>
<th>NCprec</th>
<th>SCprec</th>
</tr>
</thead>
<tbody>
<tr>
<td>TORT-CTRL</td>
<td>-5.32</td>
<td>1.04</td>
<td>76.25</td>
<td>-2.73</td>
<td>-0.56</td>
<td>0.57</td>
<td>-125.3</td>
</tr>
<tr>
<td>TORTPD-CTRL</td>
<td>-4.52</td>
<td>0.67</td>
<td>121.3</td>
<td>-1.85</td>
<td>-0.03</td>
<td>0.67</td>
<td>-79.67</td>
</tr>
<tr>
<td>PDTORT-CTRL</td>
<td>-0.57</td>
<td>0.49</td>
<td>-13.74</td>
<td>-0.34</td>
<td>-0.43</td>
<td>0.18</td>
<td>-33.76</td>
</tr>
<tr>
<td>Synergy effect*</td>
<td>-0.23</td>
<td>-0.12</td>
<td>-31.31</td>
<td>-0.54</td>
<td>-0.1</td>
<td>-0.28</td>
<td>-11.87</td>
</tr>
</tbody>
</table>

a. Synergy effect= (TORT-CTRL)-(TORTPD-CTRL)-(PDTORT-CTRL)= (TORT-TORTPD)-(PDTORT-CTRL)
Data from the whole Late Miocene period are used. In comparison, mammal fossil data have better dating control. Only the data from the early Late Miocene (i.e. the Tortonian) are used to compare with our model experiments that are designed for representing the early Late Miocene.

It has been shown in Sect. 2.5 that our models have some biases in simulating both temperature and precipitation (Fig. 1 and 2). To avoid the influence of such model biases on the agreement between model and proxy for the Late Miocene, we chose to compare the modelled anomalies between the Tortonian and present day with the anomalies between proxy and modern observation (CRUTS 2.1: http://www.cru.uea.ac.uk/cru/data/hrg/cru_ts_2.10) rather than the absolute values from model and proxy. This technique improves the agreement between model and proxy. In addition, the climate anomalies to the present day are also more relevant to palaeoclimatic studies which usually use climate anomalies rather than the absolute values to characterize the climate state of the geological past. All our modelling results are bilinearly interpolated to the corresponding fossil localities for the comparison with the proxy data.

4.6.1. Comparison with temperature reconstructions

Temperature anomalies between the Late Miocene and present day depicted by the proxy data are shown in Fig. 7a and b. Both the plant and mammal fossil data indicate generally higher mean annual temperature over the Asian continent in the Late Miocene. The increase of surface temperature is most pronounced in high-latitude regions and the TP, but is small over the low-latitudes. There are both plant and mammal fossil localities in southern China and northern India showing lower temperature in the Late Miocene (Fig. 7a and b). This suggests that cooler conditions over these regions might be a robust feature.
in the Late Miocene.

The modelled temperature anomalies of the Late Miocene at the fossil localities are in general agreement with the proxy data (Fig. 7c-f). Both the global and regional models capture the general warmer conditions over the Asian continent. However, compared to the proxy data, the increase of temperature over the high-latitude region is less prominent, and the decrease of temperature over southern China and northern India is not observed in our model experiments. The temperature anomalies indicated by the regional model are very similar to that of the global model, except that the warming in the high-latitude localities is slightly stronger and the increase of temperature over southern China and northern India is less than in the global model because of the stronger drop of winter temperature simulated in our regional model (see more discussion in Sect. 4.3). Therefore, the regional model yields a better agreement in temperature changes with the proxy data.

### 4.6.2. Comparison with precipitation reconstructions

Fig. 8 displays precipitation anomalies between the Late Miocene and the present day based on both proxy and model data. Different from temperature anomalies, the reconstructed mean annual precipitation anomalies by plant and mammal fossil data exhibit distinct patterns. Overall wetter conditions over most of the Asian continent in the Late Miocene are indicated by plant fossil. Only several plant fossil localities over southern China and northern India show negative precipitation anomalies. In contrast, widespread drier conditions over northern China and the Indian and Indochina Subcontinent in the Late Miocene are inferred by mammal fossil data. Wetter conditions are observed mainly in Central Asia.

Both the global and regional model experiments suggest a reduced precipitation over East China and an increased precipitation over the mid-latitude region in the Late Miocene. This is in more agreement with mammal fossil data. Over India, the global model simulates an enhanced precipitation in all the localities, while the regional model produces a weakened precipitation in some localities (Fig. 8e,f). It is hard to determine which results are more consistent with the proxy data, because a mosaic pattern of precipitation changes over India is manifested in both plant and mammal fossil data. This implies large uncertainties with respect to the precipitation changes over India in the Late Miocene. More complete analysis on the comparison between model and proxy data will be carried out in the future.

### 4.7. Warming world analogy

![Fig. 8. As in Fig. 7, but for mean annual precipitation changes (mm) in the Late Miocene. (a) proxy minus present-day observation (CRU TS 2.1 data set (1901-2002)) based on plant fossil data; (b) proxy minus present-day observation based on mammal fossil data; (c) GTORT minus GCTRL for plant fossil localities; (d) GTORT minus GCTRL for mammal fossil localities; (e) TORT minus CTRL for plant fossil localities; (f) TORT minus CTRL for mammal fossil localities.](image)
Since the magnitude of global warming in the Late Miocene is equivalent to that projected for the future due to anthropogenic emission of greenhouse gases, the Late Miocene climate has been regarded as a good analogue of the future global warming climate (Utescher et al 2011). Understanding the climate changes in the Late Miocene and other geological warming periods and evaluating the model performance in these periods may guide us to a better grasp of the future trend (Kutzbach & Behling 2004, Schmidt et al 2013).

To illustrate whether the Asian monsoon development in the Late Miocene and other geological warming periods as well as the future share common traits, the Asian monsoon changes in the Late Miocene, the mid-Pliocene, mid-Holocene and the future as simulated by climate models are summarized in Table 7. Although the physical boundary conditions for these periods are distinct from each other, a stronger-than-present mean ISM rainfall is observed in all these periods in the global model simulations, indicating this might be a robust pattern in the warm global climate. However, we notice that the regional climate model document a generally weaker ISM rainfall (particularly over northern India) in both the Late Miocene and the future. This suggests the necessity of using regional climate models to depict the response of the ISM rainfall to warmer global climate and its uncertainties.

Compared to the ISM, the changes of the EASM in the warming periods are less consistent. While our results suggest a weaker EASM wind and less (more) rainfall in northern (southern) China in the Late Miocene, a stronger EASM with more (less) precipitation in northern (southern) China is observed in other periods. The different changes of the EASM in our Late Miocene simulations can be attributed to the strong influence of the lower northern TP at that time which counteracts the warming effect and suppresses the EASM (see more discussion in Sect. 4.5.2). It is also possible that our one model simulation might have bias and underestimate the influence of global warming on the EASM.

As to the winter monsoon strength, there are also discrepancies in different warming periods. While our Late Miocene runs indicate a stronger winter monsoon, a weakened winter monsoon over East Asia is found in the climate simulations for the future due to the reduced land-sea thermal contrast because of the significant warming over the inner Eurasia (Hori & Ueda 2006). The winter monsoon changes in the mid-Pliocene seem to be a mixture of the Late Miocene and future changes, displaying a weakened winter monsoon over northern China but strengthened one over southern China. As discussed in Sect. 4.3, the stronger winter monsoon in the Late Miocene is again related to the lower northern and southeastern TP at that time.

On the interannual scale, we have noticed that the ISM variability in the Late Miocene share many similarities with that in the future (see more discussion in Paper III), such as the strong ENSO and ENSO-ISM teleconnection that lead to a stronger ISM variability (Table 7). A weakened variability of NAO has also been observed in both the Late Miocene and the future. This might result in a weak influence of NAO on the ISM through circumglobal teleconnection (Ding & Wang 2005). There have been few modeling studies on the interannual variability of the Asian summer monsoons in the mid-Pliocene and the mid-Holocene. However, a stronger (weaker) ENSO variability in the mid-Pliocene (Holocene) has been reported (Table 7). Given the dominant control of ENSO on the ISM variability, a stronger (weaker) ISM variability would be expected in the mid-Pliocene (Holocene).

Overall, in terms of the Asian monsoon responses, the Late Miocene and the mid-Pliocene
Table 7. Summary of the modeling studies on the Asian monsoon in the warm periods of geological past and the future. All the results are compared to the present-day or pre-industry state.

<table>
<thead>
<tr>
<th></th>
<th>Late Miocene (This study)</th>
<th>Middle Pliocene</th>
<th>Middle Holocene</th>
<th>Future (2100 AD)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Model</strong></td>
<td>ECHAM5/MPI-OM</td>
<td>Multi-model ensemble</td>
<td>Multi-model ensemble</td>
<td>Multi-model ensemble</td>
</tr>
<tr>
<td><strong>Boundary conditions</strong></td>
<td><strong>CO₂ &amp; ice sheet</strong></td>
<td>360 ppm; No ice sheet in Greenland</td>
<td>405 ppm; Great reduction of Greenland Ice sheet and removal of Western Antarctic Ice sheet (Haywood et al. 2010)</td>
<td>280 ppm; Present-day ice sheet</td>
</tr>
<tr>
<td><strong>Orbital Par.</strong></td>
<td>Present-day</td>
<td>Present-day</td>
<td>Mid-Holocene (Zheng et al. 2013)</td>
<td>Present-day</td>
</tr>
<tr>
<td><strong>Topography &amp; Vegetation</strong></td>
<td>Lower northern TP and less desert in Central Asia (Fig. 3 in Paper I)</td>
<td>Similar to the Late Miocene (Haywood et al. 2010)</td>
<td>Present-day</td>
<td>Present-day</td>
</tr>
<tr>
<td><strong>Mean climate</strong></td>
<td><strong>Temperature</strong></td>
<td>1.5 °C warmer</td>
<td>1.8-3.6°C warmer (Haywood et al. 2013)</td>
<td>No change in global mean but 2 °C warmer over central Eurasia in summer (Braconnot et al. 2007).</td>
</tr>
<tr>
<td></td>
<td><strong>EASM</strong></td>
<td>Weaker circulation, less (more) rainfall in northern (southern) China.</td>
<td>Stronger circulation, more (less) rainfall in northern (southern) China (Zhang et al. 2013).</td>
<td>Stronger circulation, more (less) increase of rainfall in northern (southern) China (Zheng et al. 2013)</td>
</tr>
<tr>
<td></td>
<td><strong>Winter monsoon</strong></td>
<td>Stronger winter monsoon over East China and India.</td>
<td>Weaker (stronger) winter monsoon over northern (southern) China (Zhang et al. 2013).</td>
<td>Weaker winter monsoon (Hori &amp; Ueda 2006)</td>
</tr>
<tr>
<td><strong>Interannual variability</strong></td>
<td><strong>ISM</strong></td>
<td>Stronger (similar) variability of circulation (rainfall).</td>
<td>Stronger ENSO variability (HadCM3) (Haywood et al. 2007)</td>
<td>Weaker ENSO variability (Zheng et al. 2008)</td>
</tr>
<tr>
<td></td>
<td><strong>Tropical forcing</strong></td>
<td>Stronger and more regular ENSO variability; stronger ENSO-ISM teleconnection</td>
<td>Stronger ENSO variability (Haywood et al. 2007)</td>
<td>Weaker ENSO variability (Zheng et al. 2008)</td>
</tr>
<tr>
<td></td>
<td><strong>Extratropical forcing</strong></td>
<td>Mean positive NAO mode; Weaker NAO variability; weaker summer NAO-ISM teleconnection; weaker winter/spring Eurasian temperature and ISM connection.</td>
<td>Mean positive NAO mode; no change in NAO variability (Gladstone et al. 2005)</td>
<td>Mean positive NAO mode; Weaker NAO variability (Paeth et al. 1999).</td>
</tr>
</tbody>
</table>
might be better analogue of the future changes than the warm periods on orbital scale (such as the mid-Holocene). The mid-Holocene warming is largely driven by insolation changes modulated by orbital parameters. Distinct from other warming periods, the mid-Holocene does not exhibit a global-wide increase of surface air temperature, but a redistribution of warmth among seasons and regions (Table 7). This might account for the unique behaviours of the Holocene monsoon climate, such as the stronger mean ISM circulation, and the weakened ENSO (probably also ISM) variability. Recent study by Liu et al (2013) also indicates that the different heating mechanism in the mid-Holocene (more lower tropospheric heating related to solar radiation changes) from that in the future (more upper tropospheric heating related to the changes in the concentration of greenhouse gases) might affect its monsoon rainfall responses. Compared to the mid-Pliocene, the Late Miocene East Asian monsoon climate seems to be more subject to the changes in orography (particularly the lower northern TP), therefore exhibits weak EASM but strong EAWM that is distinct from that in the mid-Pliocene and the future. Since the Late Miocene orography changes had small influence on the ISM, the changes of ISM in the Late Miocene might have been more influenced by global warming, thereby displays more similarity to that in the mid-Pliocene and the future.

### 4.8. Adding value and limitation of regional model experiments

The adding values and limitations of our regional climate model experiments in simulating the Late Miocene Asian monsoon have been discussed in the published papers. Here, they are summarized in Table 8.

It is found that although the large-scale monsoon circulation in the regional model is mostly consistent with its global driving model (cf. Fig. 2c,d and Fig. 5c,d in Paper I), the spatial structure of monsoon rainfall can be better depicted by the high-resolution regional model (cf. Fig. 2b and Fig. 4d in Paper I) due to its better representation of meso-scale orography (cf. Fig. 1 and Fig. 3 in Paper I) and monsoon convection (Fig. 8 in Paper I). This provides more coherent interpretation of the Late Miocene monsoon rainfall proxies from different regions. The high-resolution regional model also shows improved performance in simulating the winter monsoon (Fig 2e,f and Fig. 5a,b), which deserves more studies in the future. A unique advantage of regional climate model is that it can separate the regional and global forcings for the monsoon climate changes (see discussion in Sect. 4.5). This can hardly be obtained from the global model.

There are also several limitations of our regional model experiments. Firstly, the model

<table>
<thead>
<tr>
<th>Adding values</th>
<th>Limitations</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) Improved description of small-scale orography and monsoon convection (Paper I and II).</td>
<td>(1) Limited model domain and one-way nesting (Paper II).</td>
</tr>
<tr>
<td>(2) Improved description of regional difference in monsoon precipitation (Paper I and III).</td>
<td>(2) No sub-grid scale orography scheme (Paper II).</td>
</tr>
<tr>
<td>(3) Improved simulation of winter monsoon strength (Sect. 2.5 and 4.3 in Synopsis).</td>
<td>(3) No ocean- or vegetation-atmosphere interactions (Paper I, II and III).</td>
</tr>
<tr>
<td></td>
<td>(5) Only one-model simulations (Paper I, II and III).</td>
</tr>
</tbody>
</table>

Table 8. Adding values and limitations of regional model experiments in simulating the Late Miocene Asian monsoon climate.
does not cover all the areas that are vital for the Asian monsoon climate, such as the Subtropical High over the Western North Pacific. It is one-way nested in the global model, which does not allow the interaction between the regional and global circulation. This might lead to some biases in our regional model experiments, particularly those in Paper II. Secondly, the effect of sub-grid scale orography on atmospheric dynamics is not fully accounted for in the regional model version used in this study (Doms et al. 2011). A sub-grid scale orography scheme (Schulz 2008) has been recently implemented in the new version of COSMO-CLM, which might further improve the performance of our regional model in simulating the Asian monsoon. Thirdly, the regional model does not include the ocean-atmosphere or vegetation-atmosphere interaction, which might be important for the monsoon changes in the Late Miocene. We use relatively coarse spatial resolution in the regional model, which can hardly display the benefits of its non-hydrostatic model code. It would be interesting to run the regional model for the Late Miocene with a very high resolution over certain regions where non-hydrostatic atmospheric motions are prominent, such as the TP, and the South and East China Sea where summer monsoon is strongly influenced by typhoon. Finally, in this study, we use only one global model and one regional model to simulate the Asian monsoon in the Late Miocene. The results might be subjected to their own model bias. The ensemble simulations of the Asian monsoon in the Late Miocene with multi-global and regional models will be helpful to reduce model biases and the uncertainties of our understanding on the Asian monsoon climate in the Late Miocene.

5. Concluding remarks

In this study, we employed a relatively high-resolution regional climate model COSMO-CLM nested in a fully coupled atmosphere-ocean general circulation model ECHAM5/MPIM-OM to investigate the spatio-temporal evolution of the Asian monsoon climate in the Late Miocene and its causes. The main results and implications of this work can be summarized as follows:

1. The regional model can better characterize the regional pattern of the monsoon climate changes in the Late Miocene compared to its driving global model. It indicates that while the monsoon climate might have established or even stronger-than-present over southern China and most of India in the early Late Miocene (i.e. Tortonian), it may have not fully developed over northern China and northern Pakistan at that time. There might have been an inverse phase change of summer rainfall between northern (decrease) and southern (increase) China, which resolves the apparent discrepancies in the monsoon proxies from these two regions.

2. Both global climate and the regional tectonic changes may have exerted great impact on the monsoon changes in the Tortonian. The warmer Tortonian global climate may have induced more monsoon rainfall over India and southern China, while the lower northern and southeastern TP may have favoured a weaker summer monsoon over northern China, and a stronger winter monsoon over East China and India that led to a decrease of surface air temperature in some of the regions. The response of monsoon climate to the regional tectonic changes may be more sensitive (or even different) under the Tortonian global climate conditions than that under the present-day global climate conditions.

3. The regional mountain uplift can give rise to an asynchronous development of the Indian summer monsoon (ISM) and the East
Asian summer monsoon (EASM). While the ISM is mainly strengthened by the uplift of the southern TP (the Zagros Mountains) due to its thermal insulation (mechanical blocking) effect, the EASM is mainly enhanced by the uplift of the central and northern TP due to their diabatic heating effect. These results provide alternative explanations to the asynchronous changes of the ISM and EASM in the Neogene.

(4) The interannual variability of the ISM in the Late Miocene might have been as strong as or even stronger than today, owing to the strong ENSO and ENSO-ISM teleconnection at that time. The extratropical forcings on the interannual variability of the ISM may have been weak in the Late Miocene, due to the warmer global climate and the stabilized NAO.

(5) The regional model experiment for the Toronian displays better agreement with proxy data in temperature than its driving global model. Both the regional and global model runs show better consistency with the precipitation reconstructed by mammal fossil data than that derived from plant fossil data for the Late Micoene.

(6) Although the Late Miocene Asian monsoon climate might have been strongly influenced by regional tectonic changes (e.g. the lower northern TP), it still exhibits great resemblance to that projected for the future (particularly for the ISM). This implies similar processes for the monsoon changes may operate in both the Late Miocene and the future. Better knowledge and model simulations of the Late Miocene monsoon climate may provide useful constraints on the prediction of the future Asian monsoon changes.
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