Palaeogeography, Palaeoclimatology, Palaeoecology 365–366 (2012) 302–311

Contents lists available at SciVerse ScienceDirect

Palaeogeography, Palaeoclimatology, Palaeoecology

journal homepage: www.elsevier.com/locate/palaeo

Eocene monsoon prevalence over China: A paleobotanical perspective

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ARTICLE INFO

Article history:
Received 22 May 2012
Received in revised form 25 September 2012
Accepted 27 September 2012
Available online 4 October 2012

Keywords:
Eocene climates
Plant fossils
Quantitative reconstruction
Eocene monsoon over China

ABSTRACT

Proxy-based quantitative estimates of Eocene climates can be made from marine isotope records for ocean conditions or fossil plants for terrestrial environment. However, our understanding about Eocene terrestrial climates is derived mainly from North America and Europe, and little is known about East Asia. Previous qualitative paleoclimate studies briefly revealed three climatic regimes across China during the Eocene with a planetary wind-dominated subtropical to tropical arid zone in the central part (i.e., the subtropical highs), which was flanked by the subtropical climate zone in the north and tropical climate zone in the south. But such a pattern of paleoclimatic zonation still requires a test from quantitative study. Based on analyses of 66 plant assemblages, carefully selected from 37 fossil sites throughout China, we here report the first large-scale quantitative climatic results and discuss the Eocene climatic patterns in China. Our results demonstrate that the Eocene monsoonal climate must have been more or less developed over China, judging from the presence of apparent seasonality of both temperature and precipitation revealed by our quantitative estimation. This appears not to support the previously claimed Eocene planetary wind-dominated climate system, at least in the region of eastern China. In addition, the research indicates that, with a slight declining trend of MAT during the Eocene, the winter temperature substantially dropped in tropical southern China during the middle to late Eocene interval. This might be related to the development of a weak Eocene Kuroshio Current in the southwestern Pacific, and/or a significantly enhanced paleo-winter monsoon from Siberia.

1. Introduction

The Eocene environments, as generally understood, are characterized by distinct warm climates with overall global temperatures believed to be much higher than those in any other periods of the Cenozoic (Greenwood and Wing, 1995; Zachos et al., 2008; Huber and Caballero, 2011). The temperatures reached the highest levels in the Cenozoic during the Paleocene–Eocene Thermal Maximum and Eocene Thermal Maximum 2 (Wing et al., 2005; Zachos et al., 2008), and then decreased to a long-lived climatic optimum in the middle Eocene followed by an "ice-house" with small ephemeral ice-sheets in the early Oligocene (Zachos et al., 2008; Eldrett et al., 2009). Eocene climatic conditions have been well reported from proxies of either isotopes from surface-and-deep oceanic deposits or continental minerals and floras from both North America and Europe (Greenwood and Wing, 1995; Mosbrugger et al., 2005; Wing et al., 2005; Zachos et al., 2008; Greenwood et al., 2010; Utescher et al., 2011). However, little is quantitatively known from East Asia, except for several studies in China (e.g., He and Tao, 1997; Su et al., 2009; Yao et al., 2009; Wang et al., 2010; Quan et al., 2011, 2012). Because quantitative paleoclimatic estimates are essential to understand these long-term climate conditions, the paucity of studies from East Asia evidently prevents us from understanding and modeling the Eocene climates in a global context (Shellito and Sloan, 2006; Huber and Goldner, 2012).

As a vast country in East Asia, China preserves abundant Eocene palyno- and mega-floras (Fig. 1), but unfortunately no large-scale quantitative study has been conducted on the distribution of the Eocene climatic pattern in China. Based on data from the Chinese Eocene palynofloral assemblages, Song et al. (1983) qualitatively subdivided the Eocene climates of China into three zones, i.e., a humid warm temperate to subtropical zone in the north (Zones I in Fig. 1), an arid zone in the middle (Zone II), and a tropical to subtropical zone in the south (Zone III). This three-zone recognition was later supported by evidence from megafossil plants and lithological records (Guo, 1985; Liu, 1997; Sun and Wang, 2005). Although these studies have provided a possible general Eocene climatic pattern in China, it is still largely unclear about the details of Eocene climates over this vast region in East Asia due to the absence of large-scale quantitative reconstructions. Therefore, a critical question is to know what kind of climatic system, planetary wind or monsoon,

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0031-0182/$ – see front matter
http://dx.doi.org/10.1016/j.palaeo.2012.09.035

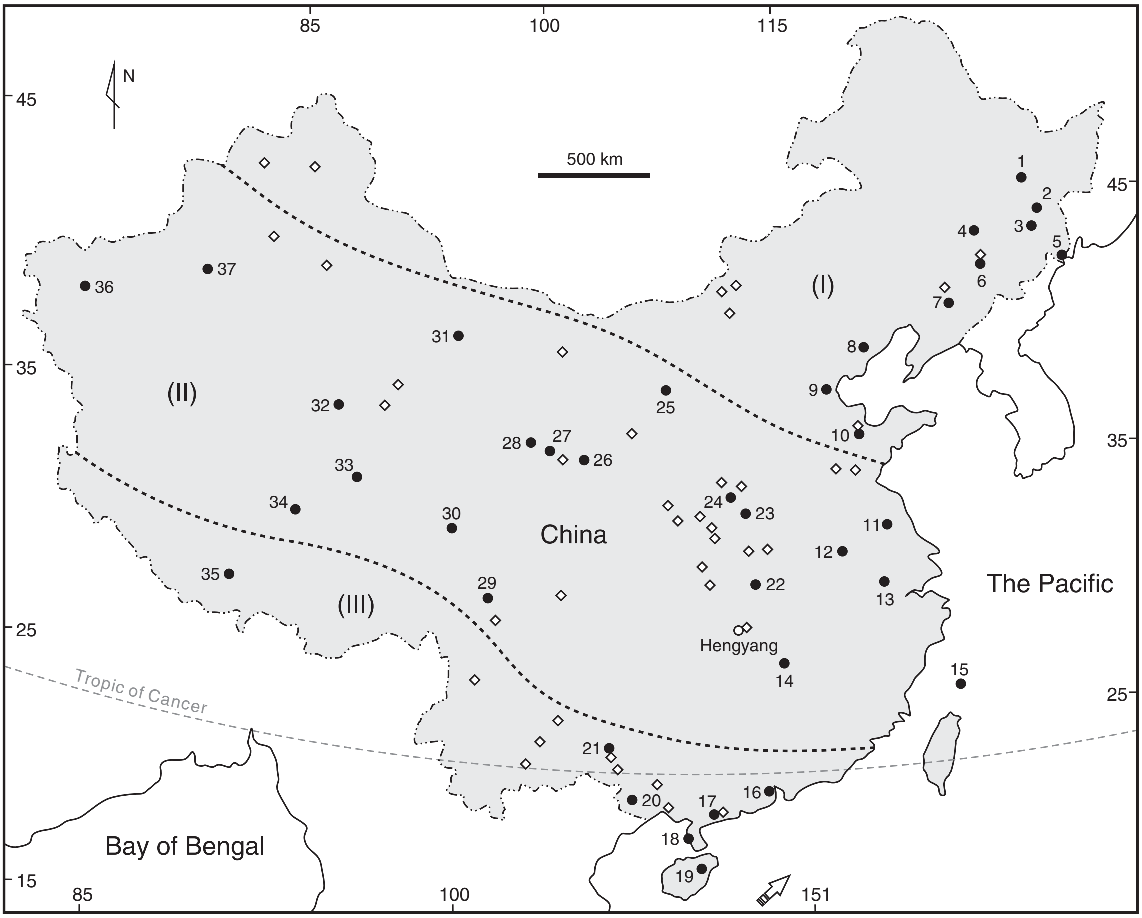

Fig. 1. Location of plant fossil sites. Each site may include several plant fossil assemblages. Site numbers as in Table 1. Climatic zone subdivision (I–III) modified from previous qualitative studies (Song et al., 1983; Guo, 1985; Liu, 1997; Sun and Wang, 2005): I—humid warm temperate to subtropical zone; II—middle arid zone (subtropical highs); III—tropical to subtropical zone. Solid circle—Eocene site yielding either mega- or microfossil plants; open diamond—Eocene mammalian fauna (data mainly from Tong, 1989). Arrow shows the Kuroshio Current.

prevailed during the Eocene, especially in central China. Moreover, the qualitative results, though important, are not always suitable for climate modeling experiments, which heavily depend upon quantitative climatic estimates (Shellito and Sloan, 2006; Huber and Goldner, 2012).

We here quantitatively reconstruct the Eocene climates by using fossil plants, including both megafossils and palynomorphs from China. Our results show that the distribution of Eocene climates is fairly even, but seasonality appears prominent, indicating a more or less developed monsoonal climate over China.

2. Methods and material

2.1. Selections and age control of fossil plant assemblages

The Eocene deposits are widely distributed throughout China, and are generally dominated by non-marine facies except the southern margin of the Tibetan Plateau. The continental strata with plenty of plant fossils are mainly fluvial, lacustrine in origin in eastern China, but in central-western China there contains red-beds and evaporites. The majority of these strata, however, lacked extensive geochronological investigations (Li, 1984). As a result, ages of many of these floras were initially assigned only to a wide geological range based on assemblages, such as to an epoch or system level. This apparently reduces the resolution of paleoclimatic results, or even casts doubt on their validity in paleoclimatic modeling. Fortunately, recent interdisciplinary studies significantly improve the age constraints in many Eocene localities (e.g., Huang et al., 1998; Wang et al., 1999; Miao et al., 2008; Shi et al., 2008; Pei et al., 2009). This provides opportunities to reconstruct paleoclimates in a better resolution at the stage level.

Although dozens of Eocene sections have been reported, 66 plant assemblages from 37 localities with well-defined ages are included in the present study (Table 1). Except for a few localities, the assemblages are selected only when at least 1 (for most cases, >2) of the following criteria well matches the age indicated by plant assemblage, i.e., 1) polarity chron studies, which were carried out in many of the localities; 2) geochemical dating, either by isotopic dating on volcanic intercalations near the fossil beds or by fission-track dating on the fossiliferous rock complexes; 3) vertebrates, especially mammals, which are main age markers in the subdivision of Chinese Cenozoic terrestrial sediments; 4) invertebrates, which are available and well studied in many localities; and 5) intensive biostratigraphical correlations between adjacent basins.

2.2. Quantitative paleoclimate reconstructions

Although several plant fossil-based methods have been developed and well used in recent decades for quantitative reconstruction of paleoclimates (e.g., Wolfe, 1979; Wing and Greenwood, 1993; Greenwood et al., 2004; Su et al., 2010; Spicer et al., 2011), the coexistence approach (CA) is applied in this study. Choice of the CA is because this approach is organ-independent and eligible to both palyno- and mega-plant fossils (Mosbrugger and Utescher, 1997; Utescher et al., 2007; Bruch et al., 2011; Utescher et al., 2011). The CA assumes that climatic tolerances of a fossil plant are not significantly different from its nearest living relatives (NLRs) when its modern affinity is determinable

Table 1
List of selected Eocene fossil sites of China. Site numbers as shown in Fig. 1. Age control except for plant assemblages (PA): C—charophyte; CN—calcareous nannofossils; D—dinoflagellate; Fo—foraminifera; G—gastropods; GD—geochemical dating; M—mammals; MA—marine animals; O—ostracods; PD—paleomagnetic dating; R—reptiles; SC—stratigraphical correlation; V—vertebrates. Age: E—early Eocene; M—middle Eocene; L—late Eocene. Fossil type: L—leaf; P—pollen and spores.

Site	Location	Coordinate	Age control	PA	Formation/group (member)	Age	Fossil type	Reference
1	Yilan, Heilongjiang	46.1°N, 129.3°E	GD	1	Xin'ancun	E	P	Quan et al. (2012)
				2	Dalianhe	M	P; L	Liu (1990); He and Tao (1997)
2	Hualin, Heilongjiang	44.8°N, 129.8°E	PD; SC	3	Bahuli	E	P	Quan et al. (2012)
3	Mudanjiang, Heilongjiang	44.6°N, 129.4°E	GD	4	Huanghua	M	P; L	Quan et al. (2012)
4	Shulan, Jilin	44.5°N, 126.9°E	M; PD	5	Bangchuigou	E	P	Fan (1985)
				6	Jishu	M	P	
5	Hunchun, Jilin	42.7°N, 130.5°E	SC	7	Hunchun (lower)	M	P	Liu, (1987); Zhang et al. (1987)
				8	Hunchun (upper)	L	P	Zhang et al. (1987)
6	Huadian, Jilin	42.9°N, 126.7°E	M	9	Huadian	M	P; L	Zhang et al. (1986); Manchester et al. (2005)
7	Fushun, Liaoning	41.8°N, 123.9°E	GD; PD	10	Guchengzi	E	P	Hong et al. (1980)
				11	Jijuntun	M	P; L	
				12	Xilutian	M	P	
				13	Gengjiajie	L	P	Qu (1993)
8	Qinhuangdao, Hebei	39.9°N, 119.6°E	GP; O	14	Kongdian (part I)	M	P	APE and NIGP (1978)
				15	Shahejie (part II)	L	P	
9	Huanghua, Hebei	38.3°N, 117.3°E	GD; PD	16	Shahejie (part IV)	E	P	Zhang and Yin (2005)
				17	Shahejie (part III)	M	P	
10	Changle, Shandong	36.7°N, 118.8°E	G; V	18	Wutu	E	P	Wang (2005)
11	Gaoyou, Jiangsu	32.8°N, 119.4°E	O	19	Dainan	E	P	Zhang and Qian (1992)
				20	Sanduo (lower)	M	P	
				21	Sanduo (upper)	L	P	
12	Hefei Basin, Anhui	31.9°N, 117.2°E	C; D	22	Dingyuan (part III)	E	P	Wang et al. (1987)
				23	Dingyuan (part IV)	M	P	
				24	Dingyuan (part V)	L	P	
13	Xuanzhou, Anhui	30.9°N, 118.7°E	D	25	Shuangta	M	P	Li (2005)
14	Qingjiang, Jiangxi	27.9°N, 116.1°E	C; M; O	26	Qingjiang (part I)	E	P	He and Sun (1977)
				27	Qingjiang (part II)	M	P	
				28	Linjiang (lower)	L	P	
15	Donghai, Zhejiang	26.4°N, 121.7°E	Fo; CN	29	Oujiang	E	P	Zhang et al. (1990)
				30	Wenzhou	M	P	
				31	Pinghu	L	P	
16	Zhujiang Basin, Guangdong	22.6°N, 113.3°E	GD; MA	32	Lufeng	E	P	Li (1998)
				33	Shenhu	M	P	
17	Maoming, Guangdong	21.5°N, 110.8°E	M; R	34	Youganwo	L	P	Yu and Wu (1983)
18	Leizhou, Guangdong	21.1°N, 109.7°E	CN; O	35	Liushagang (part II)	M	P	Zhang (1981)
				36	Liushagang (part I)	L	P	
19	Qiongshan, Hainan	19.7°N, 110.4°E	PD	37	Changchang	M	P; L	Jin et al. (2009); Yao et al. (2009)
20	Ningming, Guangxi	22.1°N, 107.1°E	M	38	Yongning Gr. (upper)	L	P	Wang et al. (2003)
21	Baise, Guangxi	23.9°N, 106.6°E	GD; M	39	Nadu	L	P; L	Guo (1979); Liu and Yang (1999)
22	Jianghan Basin, Hubei	30.4°N, 112.8°E	O	40	Xingouzui	E	P	Wang and Zhao (1980)
23	Luanchuan, Henan	33.8°N, 111.6°E	M; R	41	Tantou (lower)	E	P	Wang et al. (1984)
				42	Tantou (upper)	M	P	
24	Lingbao, Henan	34.5°N, 110.8°E	O; V	43	Xiangcheng (part IV)	E	P	Sun et al. (1985)
25	Etuoke, Inner Mongolia	39.1°N, 107.9°E	M	44	Unamed unite1 (lower)	E	P	Song and Zhang (1990)
				45	Unamed unite 1 (upper)	M	P	
26	Lanzhou, Gansu	36.1°N, 103.8°E	PD	46	Unamed unite 2 (lower)	E	P	Ma et al. (1995)
				47	Unamed unite 2 (middle)	M	P	
				48	Unamed unite 2 (upper)	L	P	
27	Minhe, Qinghai	36.2°N, 103.5°E	O; PD	49	Xining Gr.	L	P	Yu et al. (2003)
28	Xining, Qinghai	36.5°N, 101.7°E	O; PD	50	Qijiachuan (parts III, IV)	E	P	Sun et al. (1980)
				51	Honggou (parts I, II)	M	P	
				52	Honggou (parts III, IV)	L	P	
29	Litang, Sichuan	29.9°N, 100.3°E	C; GD	53	Relu (middle)	M	P	Wei and Luo (2005)
30	Shiqu, Sichuan	33.1°N, 98.6°E		54	Relu (upper)	L	L	Chen et al. (1983)
31	Yumen, Gansu	40.3°N, 97.1°E		55	Huoshaogou (Shanmacheng)	M	P	Miao et al. (2008)
				56	Huoshaogou (Qiaojia)	L	P	
32	Mangya, Qinghai	38.3°N, 90.7°E	C; PD	57	Lulehe (upper)	M	P	Zhu et al. (1985)
				58	Xiaganchaigou (lower)	L	P	
33	Kunlun Pass, Qinghai	35.3°N, 92.3°E	D; PD	59	Lulehe (upper)	M	P	Zhu et al. (1985)
				60	Wanbaogou Gr. (upper)	L	P	Guo et al. (2006)
34	Tanggula, Tibet	33.5°N, 90.2°E	G; GD	61	Totohe	M	P	Duan et al. (2007)
35	Xigaze, Tibet	29.3°N, 88.9°E	GD	62	Dagzhuka	L	P	Li et al. (2009)
36	Shache Basin, Xinjiang	38.3°N, 77.3°E	CN; Fo	63	Kalataer	E	P	Wang et al. (1986)
				64	Wulagen	M	P	Zhao et al. (1982)
				65	Bashibulake	L	P	
37	Kuche Basin, Xinjiang	41.7°N, 82.9°E	CN; D	66	Xiaokuzibai	L	P	Zhao et al. (1982)

(Mosbrugger and Utescher, 1997). For a given fossil plant assemblage, the "coexistence interval" of each climatic parameter determined by corresponding NLRs therefore represents comparable climatic tolerances of fossil counterpart, and hence environmental conditions of the given fossil location (Mosbrugger and Utescher, 1997; Pross et al., 2000; Mosbrugger et al., 2005; Utescher et al., 2007).

The data of Eocene plant assemblages analyzed here include both our collections and compilations from the literature (Table 1). Following

Liu et al. (2011), we determine NLRs of fossil taxa to a generic or familial level (Appendix A), because we cannot exactly designate a Paleogene taxon to a modern species. For NLR determinations of fossil pollen and spores, we mainly follow Song et al. (1999, 2004) and Wang (2006), who comprehensively reviewed the late Cretaceous to Neogene palynological records in China. The extinct, cosmopolitan, relict, and aquatic taxa are all excluded in the reconstruction process (Appendix A), because either they contribute little to the determination of climatic coexistence intervals, or their NLRs may no longer occupy the range of climatic conditions they can tolerate (see Hickey, 1977; Mosbrugger and Utescher, 1997; Liu et al., 2011). Additionally, to a large extent, this exclusion can avoid the main problem caused by the evolutionary gap of some taxa and the consequent autecological bias (Mosbrugger and Utescher, 1997; Pross et al., 2000; Quan et al., 2011).

By querying from the Palaeoflora Database (Utescher and Mosbrugger, 1997–2012), seven climatic parameters are reconstructed, i.e. mean annual temperature (MAT, °C), mean temperature of the coldest month (CMT, °C), mean temperature of the warmest month (WMT, °C), mean annual precipitation (MAP, mm), mean precipitation of the driest month (low month precipitation; LMP, mm), mean precipitation of the warmest month (WMP, mm), and mean precipitation of the wettest month (high month precipitation; HMP, mm). It should be mentioned that the methodology of the CA was recently challenged by Grimm and Denk (2012). However, although they have uncovered some erroneous or incomplete entries in the Palaeoflora Database, their work fails to provide any new insight concerning reliability of CA data and applicability of the CA on the fossil floral record. Because of the fact that the climatic tolerance intervals of extant taxa at a global scale are extracted from a large variety of resources that may provide heterogeneous resolution, from exact, digital point data to the interpretation of analogue maps and literature resources, such errors are largely unavoidable. It is clear that the Palaeoflora Database has been updated regularly. In the present study, the latest version of the Palaeoflora Database (Utescher and Mosbrugger, 1997–2012) is queried. To open for verification, we include the relevant datasets in Appendix B.

3. Results

Calculated coexistence intervals of 7 climatic parameters are shown in Table 2. In general, most estimated Eocene temperature intervals in China are relatively narrow (Table 2). During the early Eocene, it appears that MATs were equably distributed, ranging from 11.5–16.1 °C to 15.6–27.0 °C across China. Relatively warm winters throughout China are also indicated by CMTs that are no less than 0 °C for most sites. In summers, the temperature distribution appears fairly even, fluctuating only from 23.0–25.6 °C to 27.3–28.1 °C. Hydrologically, the precipitations were relatively high and somewhat equably distributed, with ranges of MAP from 735–1206 mm to 1183–1355 mm, HMP from 109–195 mm to 204–257 mm, LMP from 5–77 mm to 25–59 mm, and WMP from 70–143 mm to 99–188 mm (Table 2).

In the middle Eocene, the climates appear not to be considerably changed in comparison with those of the early Eocene, indicated by the equable warm temperatures with relatively high precipitations (Table 2). Specifically, the slightly higher MATs are detected in the south, north, and west of the studied region (Sites 19, 3 and 36, respectively; Fig. 1; Table 2). The estimated winter temperatures (CMTs), however, appear to show an overall decline trend from the south to the north, while the equal summer temperatures (WMTs) only ranged 22.8–25.6 °C to 27.3–28.1 °C, similar to those of the early Eocene (Table 2).

The temperatures slightly changed in central and southern China (Zones II and III), but noticeably increased in the north (Zone I) from the middle to late Eocene (Table 2). The precipitations also significantly changed in the late Eocene, shown by increasing values in the north (Zone 1) and south (Zone III), but decreasing values in the middle and eastern parts of central China (Zone II) (Table 2).

In summary, the quantitative results of the present study strongly suggest that the Eocene climates of China were warm and humid, and evenly distributed in annual temperature and precipitation, with MAT no less than 9.4 °C and MAP more than 735 mm (Table 2). The overall patterns show that in the Eocene southern China was warmer and wetter than the north, and both western and eastern China had higher MATs and MAPs than the central part. These patterns became further obvious in the late Eocene, evidenced by decreasing precipitations in central China but increasing MAP and seasonal precipitations in other surrounding regions (Table 2).

4. Discussion

4.1. Climatic distribution: planetary wind or monsoon-dominated?

The spatial distribution pattern of Eocene precipitations is intriguing (Huber and Goldner, 2012). As for the situation of the Eocene climate in China, previous qualitative studies illustrated that there was a widespread (semi-)arid zone in central China (Zone II in Fig. 1), because of the occurrence of red-beds, evaporites (Song et al., 1983; Guo, 1985; Wang et al., 1999; Zhang et al., 2007), and xerophytic representatives, such as *Ephedripites* and *Nitrariadites* pollen grains and *Palibinia* leaves (an extinct plant with narrow sclerophylls) (Song et al., 1983; Guo, 1985; Wang et al., 1999; Li and Chen, 2002; Collinson and Hooker, 2003; Dupont-Nivet et al., 2008). The presence of the (semi-)arid zone in central China was further connected to the effect of prevailing planetary wind, i.e., the subtropical highs, because it situated around 30° N of paleolatitude in the Eocene (e.g., Wang et al., 1999; Li and Chen, 2002; Zhang et al., 2007). The so-called (semi-)arid zone has also been considered to play an important role in restricting plant dispersals in East Asia during the Paleogene (Guo, 1985; Tiffney and Manchester, 2001). This justification may not stand, however, due to the occurrence and then intensified monsoon in China during the Eocene (Quan et al., 2011, 2012; Huber and Goldner, 2012; Zhang et al., 2012).

In the present study, up to 36 assemblages from 19 sites were compiled from central China, while 18 assemblages from 10 sites and 12 assemblages from 7 sites were selected from northern and southern China, respectively (Fig. 1, Table 2). It is interesting to note that the MAPs of those sites appear to be all higher than 500 mm (Table 2), the modern-day boundary between humid and arid climates (Sun and Wang, 2005). Even in central China, the so-called (semi-)arid zone recognized by qualitative studies, the lowest reconstructed MAP seems still higher than 735 mm (Assemblage 20 of Site 11; Table 2), indicating that the "arid" zone is probable not as dry as previously thought. This conclusion apparently conflicts with qualitative results based on either geological or paleobotanical evidence (e.g., Song et al., 1983; Guo, 1985; Liu, 1997; Sun and Wang, 2005). Condition in such a paradox is also observed in northwestern China during the Miocene. For example, it is widely believed that western China was arid during the Neogene, indicated by the presence of xeromorphic plant taxa and/or the development of red-beds and evaporites (Wang et al., 1999; Sun and Wang, 2005). However, the quantitative reconstructions by Liu et al. (2011) demonstrate that almost all the Miocene plant assemblages reflect MAP higher than 500 mm.

These "discrepancies" essentially concern not the numerical delimitation of aridity by the precipitation threshold between the humid and dry conditions, but the dynamic hydrologic system in paleoclimate. Principally, the arid climate is resulted from a reduced precipitation and an increased evapotranspiration (Rohli and Vega, 2008). In other words, when having a high annual evapotranspiration, even with a relatively high precipitation in a given area, it does not unavoidably imply a humid climate thereof, and vice versa. Therefore, when the outflow is unknown in the hydrologic circulation of a given region, it is difficult to suppose if the climate is arid or humid depending only on the inflow factor of precipitation (also see Wallace and Hobbs, 2006). Moreover, there are also alternative interpretations on the lithological evidence.

Table 2
Quantitative reconstructions of climatic parameters for all selected plant assemblages (PA), arranged by the geological age and climatic zonation indicated by previous qualitative studies as shown in Fig. 1. The numbers of sites, PA, and the age as in Table 1.

Age	Zone	Site	PA	Formation/group (member)	MAT	CMT	WMT	MAP	HMP	LMP	WMP
E	I	1	1	Xin'ancun	17.9–18.4	7.0–12.5	27.3–27.9	1035–1355	187–195	18–41	93–154
		2	3	Bahuli	13.6–18.4	3.7–12.5	23.6–28.1	961–1577	109–241	16–45	99–175
		4	5	Bangchuigou	15.7–16.1	6.6–7.0	25.4–25.6	1183–1355	134–153	25–41	93–141
		7	10	Guchengzi	15.2–16.1	6.6–7.0	23.6–25.6	1035–1355	134–143	25–59	93–143
		9	16	Shahejie (part IV)	16.5–18.4	6.6–12.5	27.3–28.1	1122–1298	187–195	19–24	93–175
		10	18	Wutu	11.6–16.1	1.7–7.8	22.8–25.6	735–1206	102–143	18–24	70–143
	II	11	19	Dainan	15.7–16.1	5.0–7.1	24.7–25.6	1183–1206	115–143	19–24	85–112
		12	22	Dingyuan (part III)	14.8–18.4	1.7–12.5	23.6–28.1	1122–1298	115–195	19–41	84–112
		14	27	Qingjiang (part I)	16.5–19.4	6.6–9.6	24.7–27.9	1122–1151	134–149	19–37	79–112
		15	29	Oujiang	16.8–20.8	10.6–13.3	24.7–28.1	1122–1355	204–236	19–43	94–154
		22	40	Xingouzui	16.8–18.4	10.6–12.5	24.7–25.0	897–1294	109–195	18–41	84–111
		23	41	Tantou (lower)	15.6–21.1	5.0–13.3	24.7–27.9	897–1298	109–195	18–37	84–154
		24	43	Xiangcheng (part IV)	15.2–21.4	6.6–13.9	22.8–27.9	1122–1298	134–195	19–45	93–195
		25	44	Unamed unite1 (lower)	15.6–21.1	5.0–13.3	24.7–27.9	1122–1298	115–196	19–43	99–180
		26	46	Unamed unite 2 (lower)	11.5–16.1	0.1–7.1	23.0–25.6	1122–1206	115–143	19–24	68–143
		28	50	Qijiachuan (parts III, IV)	15.6–21.1	5.0–13.3	24.7–28.1	897–1355	116–195	18–24	84–172
		36	63	Kalataer	15.6–27.0	5.0–25.0	24.7–28.1	823–1632	204–257	5–77	79–208
	III	16	32	Lufeng	15.6–21.1	5.0–13.3	24.7–28.1	1122–1355	115–195	19–59	84–154
M	I	1	2	Dalianhe	16.5–16.5	7.0–7.1	27.3–27.4	1122–1335	187–195	22–37	108–141
		3	4	Huanghua	17.9–18.3	7.0–10.2	24.7–27.7	1122–1355	115–195	19–38	84–154
		4	6	Jishu	15.6–16.1	5.0–7.1	24.7–25.6	1122–1206	115–143	19–41	84–141
		5	7	Hunchun (lower)	16.5–18.4	6.6–7.8	27.3–27.9	1096–1206	187–236	18–24	139–141
		6	9	Huadian	15.6–18.4	3.8–12.5	24.7–27.9	1194–1355	116–195	21–24	118–154
		7	11	Jijuntun	16.5–17.0	5.5–10.2	27.3–27.7	1183–1281	187–195	19–24	118–145
		7	12	Xilutian	15.6–18.4	6.6–12.5	24.7–28.1	1035–1298	134–195	18–41	99–154
		8	14	Kongdian (part I)	16.5–18.4	6.6–12.5	27.3–28.1	1122–1298	187–195	19–37	93–154
		9	17	Shahejie (part III)	15.7–18.4	6.6–7.8	23.6–28.1	1122–1206	134–195	19–24	93–143
	II	11	20	Sanduo (lower)	15.7–18.4	6.6–7.8	24.7–28.1	1122–1206	134–236	19–24	99–143
		12	23	Dingyuan (part IV)	14.8–16.1	1.7–7.8	23.0–25.6	1122–1206	115–143	19–24	70–143
		13	25	Shuangta	15.7–16.4	5.0–7.1	24.7–26.4	1122–1298	115–174	19–24	84–120
		14	27	Qingjiang (part II)	16.5–16.5	6.6–7.1	27.3–27.4	1183–1298	134–236	19–37	93–112
		15	30	Wenzhou	16.8–18.4	5.0–12.5	24.7–27.9	1183–1355	148–236	19–29	99–175
		23	42	Tantou (upper)	15.7–18.4	5.0–12.5	24.7–28.1	1096–1298	109–245	18–41	79–112
		25	45	Unamed unite 1 (upper)	13.3–18.4	1.7–12.5	23.6–28.1	897–1298	109–196	16–41	26–60
		26	47	Unamed unite 2 (middle)	11.3–16.1	1.7–7.1	22.8–25.6	735–1206	84–143	18–24	49–61
		28	51	Honggou (parts I, II)	13.3–21.1	0.1–13.3	23.6–27.9	897–1355	109–195	18–24	55–180
		30	54	Relu (middle)	11.7–16.4	6.4–7.1	22.8–26.4	1215–1298	115–172	19–29	82–112
		31	55	Huoshaogou (Shanmacheng)	11.0–21.7	1.7–15.6	22.8–28.3	735–774	102–172	18–29	70–154
		32	57	Lulehe (upper)	15.7–18.4	3.8–12.5	23.6–28.1	1096–1298	109–172	18–29	49–154
		33	59	Lulehe (upper)	15.7–18.4	3.8–12.5	23.6–28.1	1096–1355	109–172	16–29	27–172
		34	61	Totohe	13.3–16.5	1.7–7.1	23.6–27.4	1122–1355	115–172	19–24	84–154
		36	64	Wulagen	15.6–20.8	5.0–13.3	24.7–28.1	897–1298	109–195	18–45	99–180
	III	16	33	Shenhu	11.5–21.7	−1.0–14.8	23.0–28.1	735–1355	109–195	18–24	49–189
		18	35	Liushagang (part II)	11.3–23.9	13.6–16.6	23.8–27.9	1183–1613	167–293	19–75	85–180
		19	37	Changchang	17.0–20.8	12.6–13.3	26.0–27.9	1194–1520	180–236	21–24	120–141
L	I	5	8	Hunchun (upper)	17.9–18.4	7.0–12.5	27.3–28.1	1096–1520	216–236	14–24	173–175
		7	13	Gengjiajie	17.9–18.4	7.0–12.5	27.3–28.1	897–1355	187–195	18–24	99–154
		8	15	Shahejie (part II)	16.5–18.4	6.6–12.5	27.3–27.9	1122–1298	187–195	19–37	93–154
	II	11	21	Sanduo (upper)	15.6–20.8	5.0–13.3	24.7–28.1	1122–1298	115–172	19–29	82–172
		12	24	Dingyuan (part V)	11.6–16.1	1.7–7.8	22.8–25.6	735–1206	102–143	18–41	70–112
		14	28	Linjiang (lower)	16.5–19.4	6.6–9.6	27.3–27.9	1183–1298	134–195	19–37	93–112
		15	31	Pinghu	15.6–18.4	5.0–8.1	24.7–28.1	1122–1298	148–172	19–29	120–172
		26	48	Unamed unite 2 (upper)	11.6–16.5	0.1–7.1	23.0–27.4	1122–1520	115–195	19–24	49–154
		27	49	Xining Gr.	11.3–20.8	1.7–13.3	22.8–28.1	735–774	98–172	18–29	49–61
		28	52	Honggou (parts III, IV)	13.3–21.1	0.1–13.3	23.6–28.1	897–1298	109–195	18–24	49–60
		29	53	Relu (upper)	14.8–15.6	7.8–13.6	21.7–26.8	735–1215	90–153	24–41	50–142
		31	56	Huoshaogou (Qiaojia)	9.4–21.7	−0.1–15.6	18.9–28.3	735–774	73–172	18–29	49–61
		32	58	Xiaganchaigou (lower)	15.7–18.4	3.8–12.5	22.8–28.1	735–1298	84–172	18–29	49–60
		33	60	Wanbaogou Gr. (upper)	16.5–21.9	5.5–16.6	27.3–28.1	887–1355	187–237	8–24	41–154
		36	65	Bashibulake	13.3–20.8	−0.1–13.3	23.6–28.1	897–1355	109–195	18–56	55–172
		37	66	Xiaokuzibai	16.8–20.8	10.6–13.3	23.6–28.1	897–1298	109–195	18–56	49–60
	III	17	34	Youganwo	15.7–18.6	7.7–8.1	26.4–27.9	1122–1520	180–236	19–37	139–172
		18	36	Liushagang (part I)	11.3–20.8	0.1–13.3	23.8–27.9	1122–1520	167–195	19–67	68–141
		20	38	Yongning Gr. (upper)	17.2–18.4	5.0–12.5	24.7–28.1	1187–1355	216–245	19–24	139–154
		21	39	Nadu	15.7–16.1	6.6–7.1	25.4–27.4	1183–1206	225–236	19–24	120–143
		35	62	Dagzhuka	16.5–18.4	6.6–7.1	27.3–27.4	1183–1355	225–237	19–24	99–154

While oil-shale and coal are widely accepted as indicators of a humid climate, red-beds and evaporites do not necessarily refer to a high ridity (Parrish, 1998; Yechieli and Wood, 2002; Ziegler et al., 2003). In red-beds, it is ferric oxides that are responsible for the red color, typically occurred under an oxidative environment, but not closely associated with humid or arid climate (see Parrish, 1998). Evaporites are another example that once was tightly related to arid climate. However, recent studies show that the occurrence of evaporites is due to enhanced upward flux of groundwater that provides solutes from underlying strata (Yechieli and Wood, 2002), or strong precipitation seasonality with hydrological alternation between wet and dry seasons (Ziegler et al., 2003). Additionally, in previous qualitative studies, a (semi-)arid environment is determined by pollen assemblage with *Ephedripites* of over 15% (Sun and Wang, 2005), largely because comparative studies based on 313 pollen samples of Chinese Eocene evaporites show that the content of *Ephedripites* is about 15% in purple mudstone and 20% to 30% in both halite and glauberite, respectively (Tong et al., 2001). However, like those seen in modern plant physiology (Whittaker, 1975; Noble and Whalley, 1978; Poole and Miller, 1981; Bartholomeus et al., 2011), it would be difficult to distinguish physiological drought from the regional climatic aridification merely on the basis of color of rocks before associated sediments are comprehensively studied. For example, Tang et al. (2011) investigated the stepwise aridification of central Asia during the late Cenozoic based on both vegetation type and palynological diversity index from the northern Mt. Tianshan of Xinjiang in northwest China. Their results show unexpectedly high concentrations of *Ephedripites* (*Ephedra*, ~40%) during the wet period. A possible reason is that *Ephedripites* would have a rather wide ecological amplitude and climatic tolerance, and would be filled ecological niches before the expansion of its major niche competitors, including in those relatively humid conditions (Tang et al., 2011).

One the other hand, if central China was predominated by the subtropical highs caused by the sinking of hot and dry air of the Eocene Hadley Cell, it should largely be perennial arid and of low primary productivity, as seen in modern deserts and steppes at the horse latitude (Rohli and Vega, 2008). However, an independent microstratigraphic study on the lacustrine deposits of Hengyang in central China's Hunan Province shows that its Eocene climate was highly variable, especially in the aspect of precipitation (Tong, 1996). Hengyang is located in the central part of the so-called Eocene subtropical highs (Fig. 1), where both humid- and arid-climate indicators are co-occurred in the same section, such as the oil shale, red-beds, evaporites, and xerophytic *Palibinia* (Tong, 1996; Li and Chen, 2002). According to the microstratigraphic analysis by Tong (1996), the Eocene deposits alternately yield two kinds of varves, i.e., the light and dark varves, which indicate different paleoenvironments. The analysis of the samples with probe electronic microscopy shows that light varves are mainly composed of gypsum and salt rock, indicating dry climate, while the contiguous dark varves are organic rich and only silicon and phosphorus have been recognized, presenting relatively humid climate. Moreover, the continental clastic materials, an indicator of the development of runoffs, are only found in the dark varves. The microlithological evidence apparently shows that the Eocene precipitation is highly dynamic, alternating between the dry and humid periods, but not perennial dry, which largely coincides with the pattern of our plant-based results from adjacent areas (Table 2, Sites 14 and 22).

Zoogeographically, abundant Eocene mammalian faunas have been found in central China (Fig. 1). The mammals, as consumers, are at the top of the food chain. If these faunas, according to qualitative results, had lived in subtropical high regions in the Eocene, they would have lived in desert and steppe environment, typically developed under subtropical high pressure (Zhang et al., 2012). It is clear that the primary productivity of these two types of biomes would by no means support such an extensive animal community. Furthermore, the subtropical highs, if occurred in central part of eastern China, would have played a critical role of geographical barrier, which evidently would obstruct grand meridional migration of animals. However, study on faunal regionalism shows that the mammals were equably distributed during the Eocene, represented by the fact that plenty of families or even genera are found in both northern, central, and southern China (Tong et al., 1995; Qiu, 1996), and by the widely occurred *Rhombomylus–Heptodon* assemblage over China, a mammal assemblage indicating the humid climate (Tong et al., 1996). Moreover, mammal faunal correlation between China and North America with high-resolution paleomagnetic dating and corresponding carbon isotopic data shows that Asia is most probably the center of origin for many groups, such as perissodactyls, artiodactyls, and primates, which had migrated from south of central China (Hengyang) northward to North America via the Bering land bridge in the early Eocene (Bowen et al., 2002). Therefore, the mammalian fossil evidence from central China also supports a relatively free latitudinal exchange of mammalian dispersal across central China, or at least in the eastern part of central China. This is most likely due to the absence of climatic barrier in the region.

In the present study, our results of the Eocene biomes over China mainly include tropical seasonal forest, temperate forest, and shrubland (Fig. 2), largely indicating the seasonal dynamics of climate. Corresponding to this, our results also demonstrate that the annual distributions of both temperature and precipitation appear seasonally differentiated in all studied sites (Fig. 3), indicating monsoonal climate more or less prevailed in this vast region during the Eocene. The onset of the Asian monsoonal system is a topic highly debated. Sun and Wang (2005) inferred that the East Asian monsoon initiated

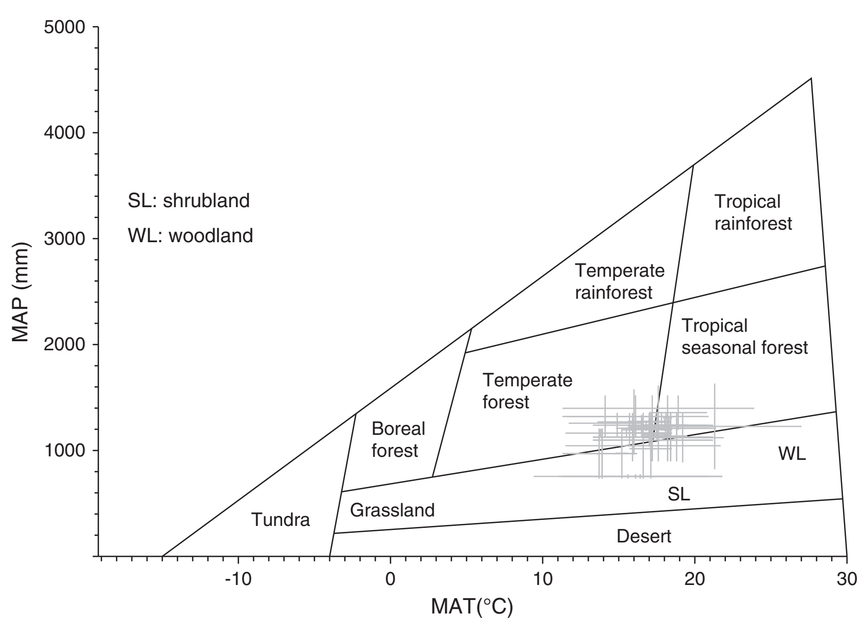

Fig. 2. Eocene climates of China illustrated by a combination of temperature and precipitation. Biomes follow Whittaker (1975).

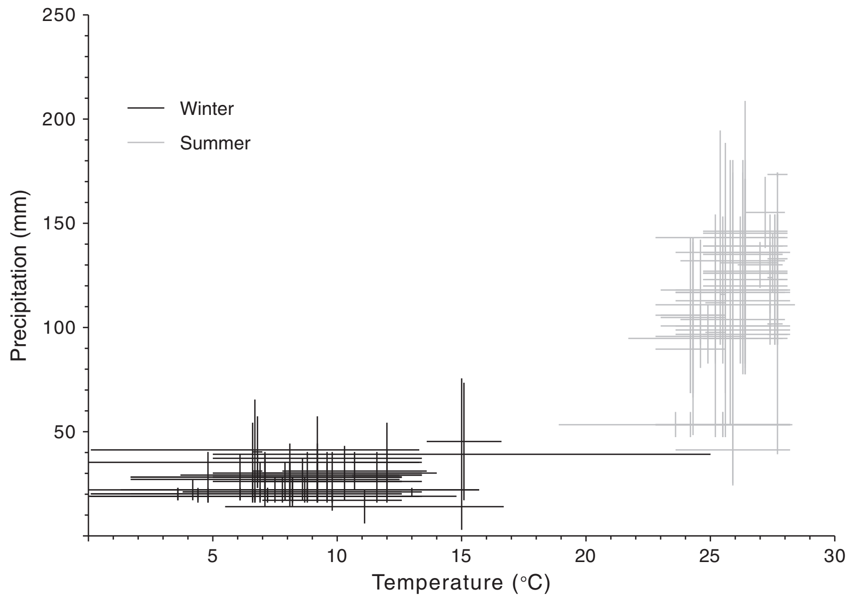

Fig. 3. Distribution of seasonal temperature and precipitation of all studied sites, showing the more or less developed monsoonal climate over Eocene China.

around the Oligocene–Miocene boundary (~23 Ma), because of the significant reorganization of paleoclimate distribution pattern in China, from a Paleogene pattern of latitudinal zonation to a Neogene pattern characterized west–east humidity contrast. However, recent quantitative studies based on plant fossils from northeast China show that the monsoon must have developed in the early Paleogene and significantly intensified in the late middle Eocene (Quan et al., 2011, 2012). The present study further reveals that the monsoonal climate appeared predominated over China during the Eocene (Fig. 3).

Our conclusion of monsoonal Eocene climate, instead of planetary wind, is consistent with recent numerical simulations. Huber and Goldner's (2012) modeling experiment, by calculating monsoonal indices and comparing them with the modern CAM3.0 Atmospheric Model Intercomparison Project simulations, indicates that there was a global-scaled monsoon system in the Eocene world, including the subsystem that dominated the majority of East Asia. Similar modeling results of monsoon are also shown by Zhang et al. (2012), but probably with a subordinate role in the Chinese Paleogene climate. Combining the modeling experiments and our quantitative reconstructions, and taking account of the uncertainties of lithological evidence, therefore, we support that the monsoonal climate must have been more or less developed in the Eocene China, at least in the majority of eastern part. The subtropical highs, if existed, would prevail at most in western China (cf. Miao et al., 2008).

4.2. Relatively stable terrestrial MAT and dramatic CMT drop in the late Eocene tropics

Although a long-term cooling climatic trend, evidenced from the warmest period in the early Eocene to the dramatic cooling Eocene–Oligocene transition when the first Cenozoic ice cap appeared on the Antarctica, has been widely reported from middle and high latitudes of both hemispheres (Wing et al., 2005; Zachos et al., 2008; Eldrett et al., 2009), the evolution of Eocene climates in the tropics is poorly known. The question of whether the Eocene tropical temperatures were relatively stable or cooling is essential to determine the polar-equatorial difference, by which the global climate is basically defined. In this section, we further focus on the evolution of Eocene

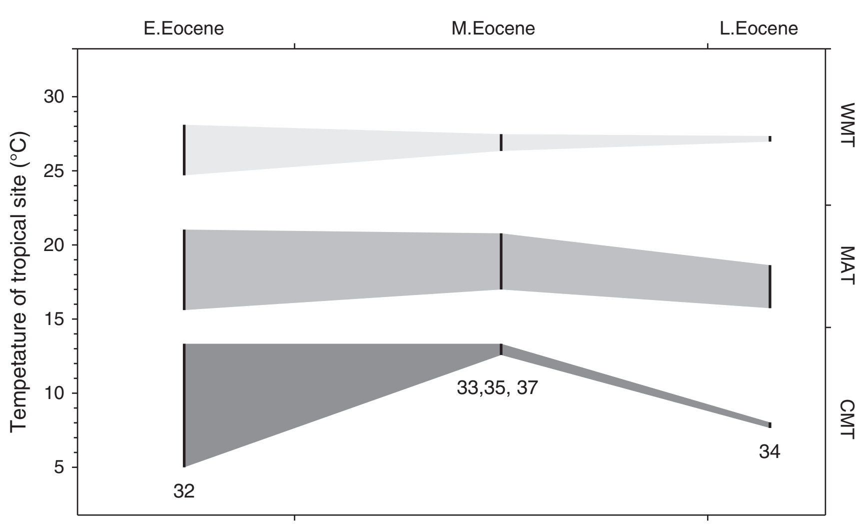

Fig. 4. Temperature evolution of the low-latitude areas. Numbers below the temperature interval bars represent the plant assemblage codes of particular stage; assemblage number as in Table 2.

terrestrial temperature in the tropics using paleobotanical data from some tropical sites in southern China (Fig. 1). The terrestrial temperatures inferred from the paleofloras are not simply equal to the SSTs of adjacent sea, but the temperatures of near-seashore lands are definitely affected by heat transported from the ocean (Sloan et al., 1995). Therefore the evolutionary pattern of inshore temperatures largely mirrors the temporal change of the SST in the overall trend (Graham, 1994).

In a total of 6 plant assemblages from 4 neighboring near-shore sites in the tropical of southern China, representing the early to late Eocene in age, were selected (Sites 16–19 in Fig. 1, Assemblages 32–37 in Tables 1 and 2). The assemblage ages are well constrained by either marine microfossils or absolute dating (Table 1). The floral data of assemblages of each geological stage were compiled together to reconstruct the overall climate conditions of this region. Among the estimated terrestrial temperature parameters, MAT slightly changed with an overall decline trend in the Eocene, from 15.6–21.1 °C, to 17.0–23.9 °C, followed by 15.7–18.6 °C (Fig. 4). However, the winter temperature dramatically decreased from 12.6–13.3 °C in the middle Eocene to 7.7–8.1 °C in the late Eocene, while the summer temperature remained almost the same with the value of 24.7–28.1 °C, 26.0–27.9 °C, and 26.4–27.9 °C respectively in the early, middle, and late Eocene (Fig. 4). Considering a potential inherent limitation of all NLR-based methods that the temperature tolerance of the fossil plant in warm episode may lie outside the envelope of its modern counterpart, we admit that the summer temperatures here may probably provide only the lower bound of the WMT.

Dramatic climate change during the Eocene has been widely reported based on marine records from both middle and high latitudes; however, the change in low latitudes appears not to be obvious. For example, Bijl et al. (2009) reported that the continuous Eocene TEX_{86} records from the southwest Pacific SST (~65° south latitude) suggest a cooling trend from ~34 °C in the early Eocene to ~21 °C by the early late Eocene. But in the tropical area, the evolution of Eocene SST has largely been debated. The Eocene marine isotopic records suggest two opposing views on the evolution of the tropical SST, which favors either a cooling trend or a relatively stable status about the Eocene SST in the tropics. Dutton et al. (2005) reported a cooling trend SSTs based on stable isotopes and Mg/Ca of foraminifera from the Ocean Drilling Program (ODP) Site 1209 in Shatsky, Rise of the Pacific. However, TEX_{86} data of surface-dwelling planktonic foraminifers from Tanzania suggest that the cooling event is not as pronounced as previously thought (Pearson et al., 2007). In the present study, our results from the tropical sites show that the MAT trend is largely similar to that of the tropical SST illustrated by Pearson et al. (2007), only slightly changed throughout the epoch (Fig. 4). However, the winter temperature remarkably dropped in the late Eocene (Fig. 4). Taking into account of the development of monsoonal climate in the vast region of China as discussed above, it appears that the slightly decreased MAT during the late Eocene would mainly be driven by cooling winters. The progressively enhanced cold air mass formed in the high-latitudinal inland must have extended southwardly in winters and resulted in a cooling as a whole, while the relatively stable summer temperatures in low latitudes might be caused by the heat transported from tropical oceans (e.g., the summer monsoon).

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.palaeo.2012.09.035.

Acknowledgments

We are grateful to Dr. P. Kershaw (Monash University), Dr. T. Jones (East Tennessee State University), and the anonymous reviewer for their helpful comments. Funding for this research was provided by the National Basic Research Program of China 2012CB822003 and 2012CB821901, NSFC 41002004 and 41172008 to C.Q., NSF EAR-0746105 to Y.S.L., and DFG and the BiK F (Loewe Initiative of the Federal State of Hessen) to T.U.

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