



Li, S., Xing, Y., Valdes, P. J., Huang, Y., Su, T., Farnsworth, A., Lunt, D. J., Tang, H., Kennedy, A. T., & Zhou, Z. (2018). Oligocene climate signals and forcings in Eurasia revealed by plant macrofossil and modelling results. *Gondwana Research*, 61, 115-127.
<https://doi.org/10.1016/j.gr.2018.04.015>

Peer reviewed version

License (if available):
CC BY-NC-ND

Link to published version (if available):
[10.1016/j.gr.2018.04.015](https://doi.org/10.1016/j.gr.2018.04.015)

[Link to publication record in Explore Bristol Research](#)
PDF-document

This is the author accepted manuscript (AAM). The final published version (version of record) is available online via Elsevier at <https://www.sciencedirect.com/science/article/pii/S1342937X18301473> . Please refer to any applicable terms of use of the publisher.

University of Bristol - Explore Bristol Research

General rights

This document is made available in accordance with publisher policies. Please cite only the published version using the reference above. Full terms of use are available:
<http://www.bristol.ac.uk/red/research-policy/pure/user-guides/ebr-terms/>

1 **Oligocene climate signals and forcings in Eurasia revealed by plant**
2 **macrofossil and modelling results**

3
4
5
6
7
8 4 Shufeng Li^{a,b}, Yaowu Xing^{a,*}, Paul J. Valdes^b, Yongjiang Huang^c, Tao Su^a, Alex Farnsworth^b,
9 Daniel J. Lunt^b, He Tang^{a,d}, Alan Kennedy^b, Zhekun Zhou^{a,c,*}

10
11
12
13 ^aKey Laboratory of Tropical Forest Ecology, Xishuangbanna Tropical Botanical Garden,
14 Chinese Academy of Sciences, Yunnan 666303, China

15
16
17
18 ^bSchool of Geographical Sciences, University of Bristol, Bristol, UK

19
20
21 ^cKey Laboratory of Biogeography and Biodiversity, Kunming Institute of Botany, Chinese
22 Academy of Sciences, Kunming 650204, China

23
24
25
26 ^dUniversity of Chinese Academy of Sciences, Beijing 100049, China

27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65
Correspondence: Yaowu Xing, tel. (fax): +86-691-8713882, e-mail: ywxing@xtbg.org.cn;

Zhekun Zhou, tel. (fax): +86-691-8713226, e-mail: zhouzk@xtbg.ac.cn.

16 **Abstract**

18 The Oligocene represents a transitional time period from a warm climate to a cooler
19 climate that is more representative of the modern day; yet, a general view of continental
20 climate pattern and forcings are still lacking. Different proxies and models show striking
21 disparities, especially in mid-high latitudes, requiring validation of Oligocene climate
22 reconstruction in order to understand the large-scale processes that drive the observed climate
23 changes. Here, we compiled 149 macrofossil floras in the mid-high latitudes of Eurasia, then
24 quantitatively reconstructed the Oligocene climate using Coexistence Approach (CA) and
25 combined previous published paleoclimate data. During the Oligocene, Eurasian mid-high

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

latitudes were mainly dominated by a humid subtropical climate. Mean annual temperature (MAT) ranged between 5.4 °C and 25.5 °C with mean annual precipitation (MAP) ranging from 338 to 2453 mm. Three regions (Europe, central Eurasia and eastern Asia) indicate different climatic regimes, with a generally warmer and wetter climate in Europe and a colder and drier climate in central Eurasia when compared to eastern Asia. No significant reorganization of climate was observed between the Early and Late Oligocene. The climate anomalies between the Oligocene and present indicate that geographic changes (e.g. retreat of the Paratethys Sea) played an important role in shaping the climate pattern of Eurasia. By comparing the fossil data to a range of different HadCM3L model simulations of the Oligocene with differing boundary conditions (e.g. CO₂ and topography), we demonstrate similar large-scale climate spatial patterns between models and fossil data, however, models simulated much higher temperature seasonality (lower simulated winter temperatures and higher simulated summer temperatures) in Eurasia. Mean annual temperature analysis indicates that simulations with 560 and 1120 ppmv CO₂ matched better with fossil data when compared to other simulations, depending on the topography. These results provide some constraints that should be considered for future paleoclimate modeling.

Keywords: Oligocene, Eurasia, paleoclimate, fossil, simulation

1. Introduction

The Oligocene (33.9–23.03 Ma) is an important epoch of the Cenozoic because it marks an early modern ‘icehouse’ climate with the formation of large ice sheets in the Antarctic (Pälike et al., 2006; Zanzani et al., 2007). Major advances in understanding the Oligocene climate have come through the use of geochemical proxies from deep-sea sediments (Zachos et al.,

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51 2001; Pälike et al., 2006; Zachos et al., 2008), fossil proxies from terrestrial sediments
52 (Mosbrugger et al., 2005; Popova et al., 2012), and paleoclimate modelling results (Ramstein
53 et al., 1997; Eldrett et al., 2009). Proxies and modelling results suggest that the Oligocene
54 climate was generally warmer and more humid than the present day (Pekar and Christie-Blick,
55 2008). However, reconstructions from a variety of proxies and models show there are many
56 uncertainties yet to be resolved especially in the higher latitudes of the Northern Hemisphere
57 (Zanazzi et al., 2007; Zachos et al., 2008; Liu et al., 2009). The inconsistencies from different
58 proxies and models thus question the validation of the Oligocene climate reconstructions and
59 constrain our understanding of Oligocene climate forcings. Therefore, a comprehensive study
60 of Oligocene climate in time and space is required to help resolve these disparities and to
61 explore the Oligocene climate forcings.

62 Abrupt cooling at the Eocene-Oligocene transition (EOT) and the forcings exerted on the
63 climate are key topics of interest (Liu et al., 2009). Modelling and isotope studies suggest that
64 Greenhouse Gas (GHG) decline may have contributed to the EOT climate cooling (Pagani et
65 al., 2005; Pearson et al., 2009). Meanwhile, geological evidence indicates that tectonic
66 modification can also have major effects on climate (Ramstein et al., 1997; Steininger and
67 Wessely, 2000). From the Oligocene to the Neogene period, significant tectonic and
68 geographic changes occurred throughout Eurasia which may have had major effects on the
69 climatic patterns throughout the continent (Seton et al., 2012). The Paratethys Sea became
70 isolated from the Tethys Sea from the EOT onwards and retreated progressively during the
71 Miocene modifying existing oceanic and atmospheric climate patterns in Europe and central
72 Eurasia (Ramstein et al., 1997; Steininger and Wessely, 2000). In eastern Asia, the uplift of
73 the Tibetan Plateau may have intensified the Asian monsoon system and may have
74 contributed to global cooling during the Cenozoic, which is more evidenced at the EOT
75 (Dupont-Nivet et al., 2007). Other geographic changes, such as opening of the Drake Passage

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

76 and Tasman Sea gateways and the glaciation of Antarctica may also have impacts on the
77 global climate system effecting the northern higher latitude area through teleconnection
78 (Scher et al., 2015). Therefore, to better understand Cenozoic climate change, it is crucial to
79 explore Oligocene climate patterns in Eurasia and identify the climate forcings behind it.

80 Fossil data can provide important information for continental paleoclimate reconstruction
81 (Mosbrugger et al., 2005). For instance, analysis of Bulgarian flora has suggested that
82 southeastern Europe experienced warm and moderately humid subtropical climatic conditions
83 during the Oligocene (Bozukov et al., 2009). In central Europe, fossil data indicate lower
84 temperature in the Oligocene compared to the warm Eocene climate (Mosbrugger et al.,
85 2005). The carpological records in Siberia and the Russian Far East show that warm and wet
86 climate conditions prevailed in this area during the Oligocene (Popova et al., 2012). In
87 northeast China, fossil data indicate a humid climate with distinct seasonality during the
88 Oligocene (Guo and Zhang, 2001; Guo et al., 2008). These reconstructions demonstrate a
89 subtropical, moderately humid climate during the Oligocene, generally above freezing in the
90 winter with low seasonality in mid-high latitudes in Eurasia (Utescher et al., 2000;
91 Mosbrugger et al., 2005; Bozukov et al., 2009; Erdei et al., 2012; Popova et al., 2012).

92 Climate modelling has been widely used to explore Paleogene climate change (Ramstein et
93 al., 1997; Eldrett et al., 2009; Huber and Caballero, 2011). The simulations show that
94 modelled Eocene greenhouse climate conditions generally yield larger temperature
95 seasonality in the higher latitudes compared to proxies such as fossils, isotopes and
96 biomarkers (Eldrett et al., 2009; Sluijs et al., 2009; Huber and Caballero, 2011). As the Earth
97 entered the Oligocene icehouse phase, coeval with global cooling, winter temperature
98 decreased dramatically in the Northern Hemisphere, leading to greater temperature
99 seasonality (Zanazzi et al., 2007; Eldrett et al., 2009). Models also show relatively warm
100 winters and high precipitation in Eurasia, suggesting a maritime type climate during the

101 Oligocene (Ramstein et al., 1997). Although these studies have improved our understanding
102 of the Oligocene climate, there has been little in the way of direct Oligocene model-proxy
103 comparison on a continental scale. Single basin analysis can only be indicative of local scale
104 processes and not necessarily continental/global scale processes. Therefore, large continental
105 model-proxy comparisons can provide a general view of the climate pattern and help us to
106 better understand the climate forcings.

107 In this study, we have compiled a database of 149 Oligocene macro-fossil floras across the
108 mid-high latitudes of Eurasia. We quantitatively reconstructed paleoclimate data using the
109 Coexistence Approach (CA, Mosbrugger and Utescher, 1997) for 114 floras, and combined
110 this with 35 published paleoclimate records (Popova et al., 2012), to explore both temporal
111 and spatial climate variability through Eurasia in the Oligocene. A comparison against a set
112 of paleoclimate model simulations with differing boundary conditions (CO₂ and
113 paleogeography) using the UK Met Office GCM (General Circulation Model), HadCM3L is
114 then presented (Valdes et al, 2017), followed by discussion of the potential driving factors
115 leading to the observed changes in climate shown in the fossil record.

117 **2. Materials and Methods**

119 *2.1 Paleoflora data*

121 The 149 Oligocene paleofloras containing their locations were compiled for the mid-high
122 latitudes of Eurasia (Appendix 1), including 114 leaf or wood floras and carpofloras in the
123 Cenozoic Angiosperm Database (Xing et al., 2016), and 35 carpofloras from Siberia and the
124 Russian Far East (Popova et al., 2012). These fossil floras are distributed in three regions:
125 Europe, central Eurasia and eastern Asia. These floras were assigned into the Early Oligocene

126 (Rupelian, 33.9–28.1 Ma) and the Late Oligocene (Chattian, 28.1–23.03 Ma) based on the
127 geological ages derived from original papers or latest revised publications.

129 2.2 Methods

130
131 The Coexistence Approach (CA, Mosbrugger and Utescher, 1997) was used to calculate
132 paleoclimate data for 114 fossil floras, and other paleoclimate data derived from 35
133 carpofloras from Siberia and the Russian Far East (Popova et al., 2012). The CA has been
134 widely applied for quantitative terrestrial climate reconstructions of the Cenozoic
135 (Mosbrugger and Utescher, 1997; Utescher et al., 2014). Uncertainty in the CA still remains
136 due to different factors, including unclear relationship between fossil and nearest living
137 relative (NLR) taxa, different distribution and climate scope between fossil and living taxa, as
138 well as uncertainty of climate threshold for NLR taxa. However, through continued updating
139 of the climate database integrated in the CA and careful consideration of the results, this
140 uncertainty can be lessened to yield comparable paleoclimate data through time and space
141 (Utescher et al., 2014). The CA uses the climatic ranges of the nearest living relatives (NLRs)
142 of all recognized taxa with known botanical affinity in a fossil flora to determine the climatic
143 range in which the majority of the flora could coexist (Mosbrugger and Utescher, 1997;
144 Utescher et al., 2014). NLRs (and associated climate parameters) were defined according to
145 the original publications and PALAEOFLORA database (<http://www.palaeoflora.de/>). The
146 paleoclimate was calculated based on generic level if applicable. For extinct genera with
147 unclear affinities we use the family level. Problematic taxa were excluded from the analysis,
148 e.g., monotypic or relict taxa such as *Cathaya*, *Metasequoia*, *Sequoia*, *Glyptostrobus*, and
149 *Comptonia*, which have been distributed in large areas in geological times compared to the
150 present day, thus could generate climate bias if included in the CA analysis (Utescher et al.,

151 2014). The program ClimStat was used to derive 7 quantitative climatic variables, i) mean
152 annual temperature (MAT, °C), ii) mean temperature of the coldest month (CMT, °C), iii)
153 mean temperature of the warmest month (WMT, °C), iv) mean annual precipitation (MAP,
154 mm), v) mean precipitation of the driest month (DryMP, mm), vi) mean precipitation of the
155 wettest month (WetMP, mm); and vii) mean precipitation of the warmest month (WarmMP,
156 mm). The WarmMP data were not used for further analysis but shown in the supplementary
157 information to provide more details of the reconstructed climate.

158 Climate anomalies between the Oligocene and present were analyzed to explore the
159 climate change and climate forcings. The observed present climate parameters were
160 calculated for each locality based on paleo-coordinate. The paleo-coordinates of fossil
161 localities were generated using the Getech Group plc. plate model (Lunt et al., 2016). Climate
162 data for each fossil site were extracted using the Bioclimatic envelope model in the R
163 package ‘Dismo’ based on paleo-coordinates for each fossil locality. The present climate data
164 were derived from the WorldClim database (<http://www.worldclim.org>, 10 minutes in
165 resolution). The mean value of 20 km data (approximately 10 minutes resolution) were
166 extracted.

167 The fossil results were then compared with results from the HadCM3L climate models
168 (Valdes et al, 2017, with a resolution of 3.75°×2.5° longitude and latitude), including some
169 new simulations and some described in previous publications (Kennedy et al., 2015; Lunt et
170 al., 2016) from Bristol Research Initiative for the Dynamic Global Environment (BRIDGE).
171 These simulations were set up with different conditions, including different CO₂ level
172 (280ppmv, 560 ppmv, 840 ppmv, and 1120 ppmv), and different paleogeographic
173 reconstructions (provided by Robertsons and Getech Group Plc.). The differences were
174 calculated by simulation data minus fossil data for each locality at the correct paleo-
175 coordinates for the Oligocene. Different regions have different numbers of fossil localities

176 thus could yield regional bias. In order to reduce this regional bias, the mean values of three
177 regions were calculated and then the mean values of the entire study area were derived from
178 these three-regional means. The modelled temperature parameters were corrected to mean sea
179 level, because most of the fossil plants are found from lacustrine or fluviatile sediments,
180 indicating these fossil floras are very likely preserved in lower altitude basins and valleys,
181 thus closer resembling sea level climate. The best matched simulations from the Rupelian and
182 the Chattian were chosen to visualize the disparity between fossil and modelling results,
183 which were determined by the lowest anomaly in mean value of MAT between model and
184 fossil data.

185

186 **3 Results**

187

188 *3.1 Overall Oligocene climate type in Eurasia*

189

190 The range of reconstructed MAT, CMT and WMT in Eurasia were 5.4–25.5 °C, -6.9–
191 21.4 °C and 18.9–29.3 °C respectively, while MAP, DryMP, WetMP and WarmMP were
192 338–2453 mm, 2–180 mm, 84–340 mm and 19–227 mm respectively (Appendix 1). The
193 reconstructed climate variables in the majority of sites in Eurasia were characterised by a
194 warm and humid summer and relatively mild or cool winter. Overall, the reconstructed
195 Oligocene climate in most of Eurasia resembled a subtropical humid climate (i.e., Cfa and
196 Cwa climate according to the Köppen climate classification; Peel et al., 2007) with distinctive
197 seasonality dominating most of the area in Eurasia.

198

199 *3.2 Temperature variability*

200

201 The reconstructed Oligocene temperatures (median values of MAT, CMT, and WMT)
1
2 202 show prominent regional spatial variability (Fig. 1 and Fig. S1). Europe was the warmest
3
4 203 region with a range of MAT between 9.4 °C and 25.5 °C, central Eurasia was the coldest area
5
6
7 204 with a range of MAT between 5.4 °C and 21.3 °C, and eastern Asia had a moderate
8
9 205 temperature with a range of MAT between 9.1 °C and 23.6 °C (Appendix 1). The warmest
10
11 206 region in Eurasia was in the southern part of Central Europe around the Paratethys Sea (Fig.
12
13 207 1). For instance, Nagysap flora in Hungary (Hably, 1989) and Trbovlje flora in Slovenia
14
15 208 (Erdei et al., 2012) with MAT of 20.6–25.0 °C and 17.2–25.5 °C respectively, were
16
17 209 significantly warmer than others (Appendix 1).
18
19
20
21

22 210 Temperature regimes between the Early Oligocene and Late Oligocene were similar, but
23
24 211 with minor differences in different regions (Fig. 1, Fig. S1 and Appendix 1). Temperatures
25
26 212 increased slightly from the Early to Late Oligocene in central Eurasia and eastern Asia (Fig.
27
28 213 1). The median MAT, CMT and WMT in central Eurasia increased from 14.1 to 15 °C, 3.2 to
29
30 214 3.9 °C, and 23.3 to 24.9 °C respectively (Fig. S1 and Appendix 1). In eastern Asia, the
31
32 215 median MAT and CMT increased from 15.1 to 15.5 °C and 3.6 to 5.8 °C respectively, and
33
34 216 WMT remained constant value of 24.9 °C (Fig. S1 and Appendix 1). However, in Europe
35
36 217 temperature decreased slightly (Fig. 1) with the median MAT, CMT and WMT decreasing
37
38 218 from 19.2 to 18.5 °C, 9.2 to 9.0 °C, and 26.4 to 25.9 °C respectively (Fig. S1 and Appendix
39
40 219 1).
41
42
43
44
45

46 220

47
48 221 Fig. 1.
49

50 222

51 223 *3.3 Precipitation variability*

52
53
54

55 224

56
57 225 The reconstructed precipitation patterns for the Oligocene (Fig. 2 and Fig. S2) indicate
58
59 226 large spatial variations. Highest MAP and WetMP occurred in Europe and the lowest
60
61
62
63
64
65

227 occurred in central Eurasia: the range of MAP in Europe was 505–2453 mm, central Eurasia
228 338–1613 mm, and eastern Asia 470–1812 mm (Appendix 1). DryMP shows that Europe was
229 slightly drier than the other two regions (Appendix 1).

230 The results demonstrate that precipitation patterns changed slightly from the Early to Late
231 Oligocene, but with spatial disparities (Fig. 2, Fig. S2 and Appendix 1). MAP and WetMP
232 decreased in Europe, but increased in both of the other regions (Fig. 2, Fig. S2 and Appendix
233 1); DryMP increased a little in Europe and central Eurasia but decreased slightly in eastern
234 Asia from the Early Oligocene to the Late Oligocene (Fig. 2, Fig. S2 and Appendix 1).

235

236 Fig. 2

237

238 *3.4 Climate anomalies between the Oligocene and present*

239

240 The Oligocene temperatures show generally warmer conditions compared to the present
241 day in the studied regions, with the exception of some sites in eastern Asia and around the
242 Paratethys Sea, which display slightly lower WMT values in the Oligocene (Figs. 3A–C).
243 Temperature anomalies in central Eurasia and summer temperature anomalies (WMT) in
244 Europe increased with distance from the Paratethys Sea (Figs. 3A–C). The largest MAT and
245 CMT anomalies were in central Eurasia, especially in the northeast of central Eurasia (Figs.
246 3A and B). The largest WMT anomalies were found in Europe especially in western and
247 central Europe (Fig. 3C). In eastern Asia, temperature anomalies increased from south to
248 north.

249

250 Fig. 3.

251

252 Our results indicate that during the Oligocene, conditions were generally much wetter than
1
2 253 the present day, but different regions show obvious spatial variability (Figs. 3D–F). Europe
3
4 254 generally had lower DryMP anomalies but higher WetMP anomalies compare to other
5
6
7 255 regions (Figs. 3E and F). In central Eurasia, regions around the Paratethys Sea show higher
8
9 256 WetMP anomalies compared to the further inland regions (Fig. 3F). DryMP anomalies in
10
11 257 central Eurasia were much higher than other regions (Fig. 3E). In eastern Asia, sites to the
12
13 258 south were drier than they are today especially in summer, while most of sites in north and in
14
15 259 central China were much wetter than present (Figs. 3D–F).
16
17
18
19 260
20
21

22 261 *3.5 Model–data comparison: different simulations*

23
24 262

25 263 *3.5.1 The Rupelian stage*

26
27 264
28
29
30

31 265 Thirty-five modelling results for the Rupelian were compared with fossil data (Fig. 4). A
32
33 266 detailed description of the modelling setup for both the Rupelian and the Chattian is included
34
35
36 267 in the Appendix 2. Our results show that the Getech Group Plc. simulations generated lower
37
38 268 MAT compared to fossil data at CO₂ levels of 560 ppmv, however at CO₂ levels of 1120
39
40 269 ppmv they show near exact agreement with the data. In contrast, Robertsons simulations
41
42 270 predicted similar MAT compared to the CA derived fossil records when assigned with CO₂
43
44 271 levels of 560 ppmv (Fig. 4). The most prominent differences between models and fossil data
45
46 272 were in CMT and WMT, indicating this suite of simulations has higher temperature
47
48 273 seasonality, suggesting that HadCM3L has a stronger seasonal cycle than indicated by the
49
50 274 fossil record (Fig. 4). Fig. 4 indicates that simulations at CO₂ levels of 560, 840 and 1120
51
52 275 ppmv better matched with fossil data for MAT and CMT, while CO₂ level at 560 ppmv
53
54 276 achieved better results for WMT. For precipitation variables, the Getech Group Plc.
55
56
57
58
59
60
61
62
63
64
65

277 simulations matched well with fossil data, while the Robertsons simulations generated
278 slightly lower values compared to fossil data especially for DryMP. Different CO₂ levels
279 show no significant impacts on precipitation variables.

280

281 Fig. 4.

282

283 3.5.2 *The Chattian stage*

284

285 Fourteen climate simulations for the Chattian were compared to the fossil data (Fig. 5 and
286 Appendix 2). Fig. 5 indicates all the simulations generated lower MAT compared to the fossil
287 data, except for one Robertsons simulation ('tedjp') with higher CO₂ level (1120 ppmv)
288 which matched well. The Robertsons models matched better with fossil results compare to
289 the Getech Group Plc. simulations when considering MAT. All the simulations show
290 significantly higher temperature seasonality (lower CMT and higher WMT) than fossil data.
291 For precipitation, all simulations were too dry. The precipitation disparities were not as
292 significant as temperature when the different scales for these variables are considered. As
293 with the Rupelian, precipitation was less sensitive than temperature to changes in CO₂.

294

295 Fig. 5.

296

297 3.6 *Model–data comparison: Spatial variability*

298

299 Determined by the mean annual temperature (MAT) differences between model and fossil
300 data, the best matched simulations 'tecqn1' (Rupelian) and 'tedjp' (Chattian) were chosen to
301 visualize the disparity between fossil and modelling results (Figs.4–5, Appendix 2). Fig. 6
302 and Fig. 7 show similar large-scale spatial patterns between the fossil reconstructions and

1 303 simulations for the Oligocene temperature variables (Europe was the warmest region, and
2 304 central Eurasia was the coldest region; the lower latitude regions generally had higher MAT
3
4 305 compared to the higher latitudes). However, there were some disparities between the
5
6
7 306 simulations and fossil data. In particular CMT and WMT indicate lower and higher values
8
9
10 307 than those of the fossil reconstructions for both the Rupelian and the Chattian. This disparity
11
12 308 was particularly evident in the higher latitudes of central Eurasia, where boreal WMT in both
13
14 309 simulations were warmer than that of fossil records. Overall, in agreement with the analysis
15
16
17 310 in the previous section on the global scale, the models yielded a larger temperature
18
19 311 seasonality than indicated by the fossil record. Precipitation spatial patterns were similar
20
21
22 312 between the fossil data and simulations (e.g, central Eurasia is the driest while eastern Europe
23
24 313 is the wettest region). However, in some regions precipitation in the simulations was lower
25
26
27 314 than that of fossil record, in particular for boreal summer precipitation in central Eurasia
28
29 315 (Figs. 6 and 7).

30
31 316

32
33
34 317 Fig. 6.

35
36 318

37
38 319 Fig. 7

39
40
41 320

42 43 321 **4. Discussion**

44
45
46 322

47 48 49 323 *4.1 A subtropical humid climate*

50
51 324

52
53
54 325 The reconstructed Oligocene climate in both the proxies and models indicates that a humid
55
56 326 subtropical climate with a distinct seasonal cycle prevailed across Eurasia, generally
57
58
59 327 consistent with most previous climate reconstructions (Utescher et al., 2000; Mosbrugger et

60
61
62
63
64
65

328 al., 2005; Uhl et al., 2007; Erdei et al., 2012; Popova et al., 2012; Quan et al., 2012). Many
329 fossil floras show the majority elements are evergreen and deciduous taxa (Kvacek and
330 Walther, 2001; Erdei and Bruch, 2004; Erdei et al., 2012), as well as mesophytic forest mixed
331 with conifers (Akhmetiev et al., 2009; Erdei et al., 2012) in mid-high latitudes of Eurasia,
332 implying a warm-temperate or subtropical humid climate with distinct seasons existed in
333 these regions (Kvacek and Walther, 2001; Erdei and Bruch, 2004; Erdei et al., 2012). Some
334 floras contain high proportion of thermophilic taxa or deciduous elements, mostly from
335 higher latitudes of Eurasia, especially in Siberia and the Russian Far East (Kvacek and
336 Walther, 2001; Akhmetiev et al., 2009; Bozukov et al., 2009; Erdei et al., 2012). Fossil
337 evidence from eastern Asia indicates that temperate or warm temperate elements mainly
338 occurred in this region, as well as many subtropical elements such as Rutaceae,
339 Euphorbiaceae, Sapindaceae, Myrsinaceae, Apocynaceae and Boraginaceae, indicating a
340 temperate to warm-temperate climate (Tanai and Uemura, 1991; Guo and Zhang, 2001;
341 Popova et al., 2012; Quan et al., 2012). It should be noted that the humid subtropical climate
342 is a definition including Cfa and Cwa (the coldest month average temperature above 0 °C and
343 warmest average temperature above 22 °C) according to Köppen climate classification (Peel
344 et al., 2007; also refer to https://www.weather.gov/jetstream/climate_max). Therefore, the
345 temperate to warm-temperate climate discussed above can broadly be included in Cfa and
346 Cwa, i.e., humid subtropical climate.

347

4.2 Climate shift from the Early to Late Oligocene

349

350 Different studies have shown inconsistent results for climate changes from the Early to
351 Late Oligocene (Erdei and Bruch, 2004; Mosbrugger et al., 2005; Popova et al., 2012; Quan
352 et al., 2012). The macro-floras in Hungary indicate that the Late Oligocene had lower MAT

1 353 and CMT compared to the Early Oligocene (Erdei and Bruch, 2004). In central Europe, fossil
2 354 data reveal significant warming during the uppermost part of the Late Oligocene, coinciding
3
4 355 with the Late Oligocene warming events (Mosbrugger et al., 2005). In western Siberia, WMT
5
6
7 356 increased slightly from the Early to Late Oligocene, but CMT decreased a little (Popova et al.,
8
9
10 357 2012). While in China, the leaf fossil and pollen data indicate a cooling trend during the Late
11
12 358 Oligocene, which may be linked to Tibetan uplift and monsoon evolution (Quan et al., 2012).
13
14 359 Taken together with this evidence, our results suggest that the differences between the Early
15
16
17 360 and Late Oligocene show strong spatial heterogeneity patterns. Modelling results support the
18
19 361 fossil data, suggesting that there is little change in global climate between the Early and Late
20
21
22 362 Oligocene. However, the modelling data do display diverse spatial variability (Fig. 6 and Fig.
23
24 363 7), likely a result of perturbations in geographic and orographic forcing through the
25
26
27 364 Oligocene. Overall, the fossil data suggest that any signals are indicative of local/regional
28
29 365 change and not that of a global signal, corroborated by our suite of GCM simulations of the
30
31
32 366 Early and Late Oligocene.

33
34 367

35 36 368 *4.3 The influence of paleogeography on climate patterns*

37
38
39 369

40
41 370 Spatial patterns presented in the data for Europe and central Eurasia imply that the retreat
42
43
44 371 of the Paratethys Sea may have played an important role in the Oligocene climate regime.
45
46 372 During most of the Oligocene, the Paratethys Sea covered large areas of the central Eurasia,
47
48
49 373 thus the asymmetric heating of land and sea may lead to the warmer and humid winter
50
51 374 climate in central Eurasia (Ramstein et al., 1997). With the retreat of the Paratethys Sea, the
52
53
54 375 climate of central Eurasia became colder and dryer resulting in a typical continental climate
55
56 376 which is consistent with modelling results (Ramstein et al., 1997). This highlights the major
57
58 377 role of the Paratethys Sea as a thermal regulator of the Oligocene climate in Eurasia. The
59
60
61
62
63
64
65

1 378 Paratethys is able to facilitate the northerly advancement of sub-tropical warm moist air
2 379 masses into central Asia during the boreal summer (Fig. 8B) allowing a humid subtropical
3
4 380 climate to be sustained in more northerly latitudes that would have otherwise been located
5
6
7 381 further south.

8
9 382 Fig. 8 shows the simulated wind intensity (m/s) and trajectories at 850 hPa during the
10
11 383 Oligocene. During the Rupelian, the results indicate that westerlies prevailed in winter at
12
13 384 mid-high latitudes (Fig. 8A). In Asia in the boreal winter the lower elevations of Tibetan and
14
15 385 Mongolian plateaus compared to the present day favoured the generation of more intense
16
17 386 westerly winds over the Eurasian continent allowing greater moisture supply into central
18
19 387 Eurasia (Fig. 8A). The central Asia high-pressure system was weaker than present and may
20
21 388 have been shifted south to eastern China (Ramstein et al., 1997; Licht et al., 2016). Therefore,
22
23 389 winter was much warmer and more humid in central Eurasia during the Oligocene. Recent
24
25 390 meteorological data show remarkable weakening of the Siberian High in recent years, leading
26
27 391 to warm winter and higher mean annual temperature in higher latitudes of Eurasia, further
28
29 392 supporting our results (Gong and Ho, 2002).
30
31
32
33
34
35

36 393

37
38
39 394 Fig. 8
40

41 395

42
43 396 The atmospheric circulation changes in Eurasia not only influenced the climate of central
44
45 397 Eurasia but also influenced the monsoon system in eastern Asia (Zhang et al., 2007; Spicer,
46
47 398 2017). Previous studies based on various proxies suggest that the summer monsoon in East
48
49 399 Asia intensified through the Cenozoic due to the uplift of Tibetan Plateau and other plate
50
51 400 motions (Ramstein et al., 1997; Sun et al., 2010). Numerical modelling demonstrates
52
53 401 southwest prevailing wind intensified in summer, caused by the Tibetan Plateau uplift and the
54
55 402 Paratethys retreat, providing large amount of water and increased precipitation in the
56
57
58
59
60
61
62
63
64
65

1 403 monsoon areas in eastern Asia (Zhang et al., 2007). Sensitivity experiments also suggest that
2 404 when the Qinghai–Tibetan Plateau (QTP) is set flat (0 m), the South and Southeast Asian
3
4 405 monsoons still prevail (Liu et al., 2015). Some other fossil data also suggest that the Asian
5
6 406 monsoon established during the early Cenozoic period (Spicer et al., 2016, 2017). Our results,
7
8
9 407 combined with fossil and model data, indicate distinct precipitation seasonality in eastern
10
11 408 Asia, suggesting that monsoonal climate existed in this region. However, our results show
12
13
14 409 that summer precipitation during the Oligocene is lower than the present day in southern parts
15
16 410 of eastern Asia, which currently is remarkably affected by the Asian summer monsoon,
17
18
19 411 indicating that the summer monsoon in these regions may not have been as intense as the
20
21
22 412 present day during the Oligocene.

23
24 413 The winter monsoon in eastern Asia was dramatically different in the Oligocene compared
25
26 414 to the present day (Sun et al., 2010), and consequently influenced winter precipitation (Sun et
27
28 415 al., 2010). The winter monsoon which is dynamically linked to the strength and position of
29
30 416 the high pressure centre over the Siberian–Mongolian region normally yields a dry winter
31
32 417 (Wu and Chan, 1997). Therefore, the higher precipitation in winter may indicate lower winter
33
34 418 monsoon strength in east Asia. As discussed above, the central Asia high-pressure system
35
36 419 was weaker than present, thus can result in a weaker winter monsoon in eastern Asia (Zhang
37
38 420 et al., 2007). In addition, penetration of moisture from the Paratethys Sea into central Asia
39
40 421 also could lead to a more humid winter. This is consistent with our results that the Oligocene
41
42 422 was much wetter than today during winter, indicating a weaker winter monsoon in
43
44 423 northeastern Asia (Zhang et al., 2007; Clift et al., 2008; Sun et al., 2010).

45
46 424
47
48
49 425 *4.4 Fossil results versus models*
50
51
52
53
54
55
56 426
57
58
59
60
61
62
63
64
65

1 427 Reconstructed CO₂ proxy data can yield a large range in CO₂ concentrations during the
2
3 428 Oligocene (Beerling and Royer, 2011; Pagani et al., 2005), but the majority of data suggests
4
5 429 that atmospheric CO₂ concentrations significantly dropped during the Oligocene, reaching
6
7 430 moderate-to-low levels (280–840 ppmv) compared to the pre-industrial era (Beerling and
8
9
10 431 Royer, 2011; Pagani et al., 2005; Roth-Nebelsick et al., 2014; Zhang et al., 2013). Our results
11
12 432 show sensitivity to CO₂ concentrations when considering simulated temperature comparisons
13
14 433 with fossil data. CO₂ levels of 560 ppmv and 1120 ppmv matched better with the Rupelian
15
16 434 fossil results for Robertsons and Getech Group Plc. simulations respectively, while for the
17
18
19 435 Chattian, only one Robertsons simulation with 1120 ppmv was comparable with the fossil
20
21
22 436 data. These simulations are broadly consistent with published CO₂ results, especially for
23
24 437 Robertsons simulations of the Rupelian period. Getech Group Plc. simulations generally need
25
26
27 438 higher CO₂ level to match better with fossil data, indicating there could potentially be some
28
29 439 orographic or model spin-up impacts on simulated temperatures which need consider further
30
31 440 exploration.

32
33
34 441 The mean values from all the simulations broadly matched well with fossil data, mainly
35
36 442 because mean values generally reduced the range in the data. Consequently, multi-simulation
37
38
39 443 means with perturbed boundary conditions can produce robust results compared to a single
40
41 444 model (Knutti and Sedláček, 2013.). Nevertheless, substantial uncertainty remains in the use
42
43
44 445 of multiple simulation means, in particularly on a process-based level. Quantitative
45
46 446 evaluation of this uncertainty should be considered in further work to understand the
47
48
49 447 differences between simulations.

50
51 448 Both fossil data and climate models demonstrate a similar spatial pattern on a broad
52
53 449 continental scale, with central Eurasia being the coldest and driest region, and Europe and
54
55
56 450 eastern Asia being much warmer and more humid, implying these two results are consistent
57
58 451 and comparable. However, there are distinct differences between the models and fossil data;
59
60
61
62
63
64
65

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

452 in particular, there is prominent disparity between both fossil and modeled seasonality during
453 the Oligocene with the latter indicating a much greater seasonal range in temperature. These
454 differences can be derived from the uncertainties from both fossil reconstruction and the
455 modeling approach. The CA normally represents the climate of a particular fossil flora,
456 however, it assumes that fossil taxa has a similar climate envelope as their NLRs. Thus, there
457 is potential to generate large errors when applied to early geological ages (Utescher et al.,
458 2014). Fossil taxa may have different climate scope (or thresholds) compared to modern taxa
459 because of evolution, thus the bias may not be the same for different climate parameters. For
460 example, some taxa may be widely distributed in the geological past but now are restricted to
461 specific areas, therefore the range from CMT and/or WMT may be different than in the
462 present day, yet the equivalent MAT can still be similar to the present day. In addition, the
463 unclear identification of some NLR taxa and modern taxa climate data could also contribute
464 to uncertainties (Utescher et al., 2014).

31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

465 Climate models can produce uncertainties due to the prescribed parameters and various
466 data inputs, some of which are poorly constrained and thus they could have large effects on
467 the paleoclimate modelling results (Valdes, 2000). There have been a number of studies
468 which have compared fossil, isotope and biomarker results to those of model simulations,
469 indicating models could yield significantly lower winter temperatures and produce higher
470 temperature seasonality at higher latitudes, during past warm periods such as the Eocene and
471 Cretaceous (Herman and Spicer, 1996; Eldrett et al., 2009; Sluijs et al., 2009; Huber and
472 Caballero, 2011). CO₂ concentration is the most important and unfortunately unconstrained
473 factor that can dramatically altered the modeling results (Anagnostou et al., 2016), although it
474 has less effect on seasonality (Figs. 4 and 5). As is shown in the modeling results, when set
475 up with mid-high CO₂ levels (560–1120 ppmv) for the Rupelian, models give a good match
476 the fossil data for MAT, but the temperature seasonality remains consistent regardless of CO₂

1 477 level (Figs. 4 and 5). However, atmospheric CO₂ reconstructions for the Oligocene are
2 478 limited and different proxies display large uncertainties (Pearson et al., 2009; Zhang et al.,
3
4 479 2013), thus limiting the validation of modelling results. Geographic and orographic features
5
6
7 480 may dramatically affect the regional climate by altering the atmospheric and oceanic
8
9
10 481 circulation and the land surface-atmosphere interaction (Clark et al., 2001). Therefore,
11
12 482 different paleogeographic and orographic settings can generate differing sensitivities as was
13
14 483 highlighted in our results section, comparing Getech Group Plc. and Robertsons simulations.
15
16
17 484 Climate-vegetation interactions could also contribute to the disparities. The climate model
18
19 485 interactively predicts vegetation cover so does include vegetation climate feedbacks but may
20
21
22 486 not represent them well. Models at the Eocene/Oligocene boundary suggest that temperature
23
24 487 and the Antarctic ice sheet are highly sensitive to vegetation types (Liakka et al., 2014). Some
25
26 488 of the simulations also included alternative Antarctic ice sheet configurations but the climatic
27
28
29 489 differences in the Northern Hemisphere were small and did not change the model data
30
31
32 490 comparison suggesting an insensitivity to interhemispheric teleconnections resulting from
33
34 491 different ice sheet configurations on the vegetation and climate of Eurasia in both proxies and
35
36 492 models. Other important factors such as the ozone layers, clouds and soil data can also
37
38
39 493 contribute some uncertainties (Sluijs et al., 2009).

40
41 494 Overall, it is hard to say whether the CA results from fossil floras are able to adequately
42
43
44 495 resolve the climate variability which can lead to extremes in climate that could be recorded in
45
46 496 the fossil data. Interannual, interdecadal and/or intercentennial variability could bias the
47
48
49 497 result and comparison. However, when considering other proxies and model results (e.g.,
50
51 498 Herman and Spicer, 1996; Eldrett et al., 2009; Sluijs et al., 2009; Huber and Caballero, 2011),
52
53 499 we suggest that some parameters or mechanisms embedded in the models may profoundly
54
55
56 500 increase higher latitude temperature seasonality in Eurasia. Possible factors responsible for
57
58 501 this include: ground cover (different vegetation), clouds, topographic representation, ocean
59
60
61
62
63
64
65

1 502 circulation, equator-polar energy transport, and even different resolution of the model.

2 503 Refined data and sensitivity experiments should be considered for further studies, and the

3
4 504 feedbacks between climate and key factors may provide clues for future climate predictions.

5
6
7 505

8 9 10 506 **5. Conclusion**

11
12 507

13
14
15 508 Our study based on a large dataset of fossil floras provides a comprehensive picture of the

16
17 509 continental climate during the Oligocene. The reconstructed paleoclimates indicate a typical

18
19
20 510 humid subtropical climate with distinct seasonality during the Oligocene. Our results show a

21
22 511 spatial heterogeneity of climate change patterns between the Early and Late Oligocene. The

23
24
25 512 anomalies of the Oligocene compared to the present suggest that the plate motion and land-

26
27 513 sea distribution caused by the Paratethys Sea retreat and Tibetan Plateau uplift may have

28
29
30 514 played important role in shaping the climate of Eurasia since at least the Oligocene. We

31
32 515 compared the fossil data with a range of different HadCM3L simulations of the Oligocene

33
34
35 516 with differing boundary conditions. Our results suggest that fossil and modelling results are

36
37 517 generally spatial consistent but have some distinct disparities. Our findings, in agreement

38
39
40 518 with other studies from the Eocene and Oligocene, suggest that models generate higher

41
42 519 temperature seasonality and lower precipitation than fossil proxy reconstructions. Further

43
44
45 520 analysis indicates that different CO₂ concentration and topographic representations of Eurasia

46
47 521 may be responsible for the differences in the model-data comparisons. Middle to high CO₂

48
49
50 522 levels (560 and 1120 ppmv) and Robertsons simulations matched better with reconstructed

51
52 523 temperatures from the fossil record than Getech Group Plc. simulations. Therefore, we

53
54 524 suggest further sensitivity experiments should be conducted to explain the disparities between

55
56
57 525 model and proxies in higher latitude regions.

58
59 526

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

527 **Acknowledgements**

528

529 This study was supported by the NSFC-RCUK NERC joint project (No. 41661134049), the
530 grant of Natural Environment Research Council (No. NE/P013805/1), National Natural
531 Science Foundation of China (No. 41772026, 41372035, and U1502231), the Pioneer
532 Hundred Talents Program of the Chinese Academy of Sciences (No. 2016-062 to Y.W. Xing),
533 the Foundation of the State Key Laboratory of Paleobiology and Stratigraphy, Nanjing
534 Institute of Geology and Paleontology, Chinese Academy of Sciences (No. 153107), the CAS
535 135 program (XTBG-F01), NERC grant (No. NE/K014757/1; Cretaceous-Paleocene-Eocene:
536 Exploring Climate and Climate Sensitivity) and NERC grant (NE/L002434/1). We thank Dr.
537 Torsten Utescher and another anonymous reviewer for their many insightful comments and
538 constructive suggestions. This work is part of the NECLIME (Neogene Climate Evolution of
539 Eurasia) network.

540

541 **Appendices**

542

543 **References**

544

545 Akhmetiev, M., Walther, H., Kvaček, Z., 2009. Mid-latitude Palaeogene floras of Eurasia
546 bound to volcanic settings and palaeoclimatic events—experience obtained from the
547 Far East of Russia (Sikhote-Alin') and Central Europe (Bohemian Massif). *Acta*
548 *Musei Nationalis Pragae, Series B Historia Naturalis* 65, 61-129.

549 Anagnostou, E., John, E.H., Edgar, K.M., Foster, G.L., Ridgwell, A., Inglis, G.N., Pancost,
550 R.D., Lunt, D.J., Pearson, P.N., 2016. Changing atmospheric CO₂ concentration was
551 the primary driver of early Cenozoic climate. *Nature* 533, 380-384.

- 1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65
- 552 Beerling, D.J., Royer, D.L., 2011. Convergent Cenozoic CO₂ history. *Nature Geoscience* 4,
553 418.
- 554 Bozukov, V., Utescher, T., Ivanov, D., 2009. Late Eocene to early Miocene climate and
555 vegetation of Bulgaria. *Review of Palaeobotany and Palynology* 153, 360-374.
- 556 Clark, D.B., Xue, Y., Harding, R.J., Valdes, P.J., 2001. Modeling the impact of land surface
557 degradation on the climate of tropical North Africa. *Journal of Climate* 14, 1809-1822.
- 558 Clift, P.D., Hodges, K.V., Heslop, D., Hannigan, R., Van Long, H., Calves, G., 2008.
559 Correlation of Himalayan exhumation rates and Asian monsoon intensity. *Nature*
560 *Geoscience* 1, 875-880.
- 561 Dupont-Nivet, G., Krijgsman, W., Langereis, C.G., Abels, H.A., Dai, S., Fang, X.M., 2007.
562 Tibetan plateau aridification linked to global cooling at the Eocene-Oligocene
563 transition. *Nature* 445, 635-638.
- 564 Eldrett, J.S., Greenwood, D.R., Harding, I.C., Huber, M., 2009. Increased seasonality through
565 the Eocene to Oligocene transition in northern high latitudes. *Nature* 459, 969-973.
- 566 Erdei, B., Bruch, A., 2004. A climate analysis of Late Oligocene (Egerian) macrofloras from
567 Hungary. *Studia Botanica Hungarica* 34, 5-23.
- 568 Erdei, B., Utescher, T., Hably, L., Tamas, J., Roth-Nebelsick, A., Grein, M., 2012. Early
569 Oligocene continental climate of the Palaeogene Basin (Hungary and Slovenia) and
570 the surrounding area. *Turkish Journal of Earth Sciences* 21, 153-186.
- 571 Gong, D.Y., Ho, C.H., 2002. The Siberian High and climate change over middle to high
572 latitude Asia. *Theoretical and Applied Climatology* 72, 1-9.
- 573 Guo, S.Y., Zhang, G.F., 2001. Oligocene Sanhe flora in Longjing county of Jilin, northeast
574 China. *Acta Palaeontologica Sinica* 41, 193-210.

- 1 575 Guo, Z.T., Sun, B., Zhang, Z.S., Peng, S.Z., Xiao, G.Q., Ge, J.Y., Hao, Q.Z., Qiao, Y.S.,
2 576 Liang, M.Y., Liu, J.F., 2008. A major reorganization of Asian climate by the early
3
4 577 Miocene. *Climate of the Past* 4, 153-174.
5
6
7 578 Hably, L., 1989. The Oligocene flora of Nagysap. *Fragm. Mineral. Palaeont* 14, 83-99.
8
9
10 579 Herman, A.B., Spicer, R.A., 1996. Palaeobotanical evidence for a warm Cretaceous Arctic
11
12 580 Ocean. *Nature* 380, 330-333.
13
14
15 581 Huber, M., Caballero, R., 2011. The early Eocene equable climate problem revisited. *Climate*
16
17 582 *of the Past* 7, 603.
18
19 583 Kennedy, A., Farnsworth, A., Lunt, D., Lear, C.H., Markwick, P., 2015. Atmospheric and
20
21 584 oceanic impacts of Antarctic glaciation across the Eocene–Oligocene transition. *Phil.*
22
23 585 *Trans. R. Soc. A* 373, 20140419.
24
25
26 586 Knutti, R., Sedláček, J., 2013. Robustness and uncertainties in the new CMIP5 climate model
27
28 587 projections. *Nature Climate Change* 3, 369.
29
30
31 588 Kvacek, Z., Walther, H., 2001. The Oligocene of Central Europe and the development of
32
33 589 forest vegetation in space and time based on megafossils. *Palaeontographica*
34
35 590 *Abteilung B*, 125-148.
36
37
38
39 591 Liakka, J., Colleoni, F., Ahrens, B., Hickler, T., 2014. The impact of climate - vegetation
40
41 592 interactions on the onset of the Antarctic ice sheet. *Geophysical Research Letters* 41,
42
43 593 1269-1276.
44
45
46 594 Licht, A., Dupont-Nivet, G., Pullen, A., Kapp, P., Abels, H., Lai, Z., Guo, Z., Abell, J.,
47
48 595 Giesler, D., 2016. Resilience of the Asian atmospheric circulation shown by
49
50 596 Paleogene dust provenance. *Nature communications* 7, 12390.
51
52
53
54 597 Liu, Z.H., Pagani, M., Zinniker, D., DeConto, R., Huber, M., Brinkhuis, H., Shah, S.R.,
55
56 598 Leckie, R.M., Pearson, A., 2009. Global cooling during the Eocene-Oligocene climate
57
58 599 transition. *Science* 323, 1187-1190.
59
60
61
62
63
64
65

- 600 Lunt, D.J., Foster, G.L., O'Brien, C.L., Pancost, R.D., Robinson, S.A., 2016.
1
2 601 Palaeogeographic controls on climate and proxy interpretation. *Climate of the Past* 12,
3
4 602 1181.
5
6
7 603 Mosbrugger, V., Utescher, T., 1997. The coexistence approach—a method for quantitative
8
9 604 reconstructions of Tertiary terrestrial palaeoclimate data using plant fossils.
10
11 605 Palaeogeography, Palaeoclimatology, Palaeoecology 134, 61-86.
12
13
14 606 Mosbrugger, V., Utescher, T., Dilcher, D.L., 2005. Cenozoic continental climatic evolution
15
16 607 of Central Europe. *Proc Natl Acad Sci U S A* 102, 14964.
17
18
19 608 Pagani, M., Zachos, J.C., Freeman, K.H., Tipler, B., Bohaty, S., 2005. Marked decline in
20
21 609 atmospheric carbon dioxide concentrations during the Paleogene. *Science* 309, 600-
22
23 610 603.
24
25
26 611 Pälike, H., Norris, R.D., Herrle, J.O., Wilson, P.A., Coxall, H.K., Lear, C.H., Shackleton,
27
28 612 N.J., Tripathi, A.K., Wade, B.S., 2006. The heartbeat of the Oligocene climate system.
29
30 613 *science* 314, 1894-1898.
31
32
33
34 614 Pearson, P.N., Foster, G.L., Wade, B.S., 2009. Atmospheric carbon dioxide through the
35
36 615 Eocene–Oligocene climate transition. *Nature* 461, 1110-1113.
37
38
39 616 Peel, M.C., Finlayson, B.L., McMahon, T.A., 2007. Updated world map of the Köppen-
40
41 617 Geiger climate classification. *Hydrology and Earth System Sciences Discussions* 4,
42
43 618 439-473.
44
45
46 619 Pekar, S.F., Christie-Blick, N., 2008. Resolving apparent conflicts between oceanographic
47
48 620 and Antarctic climate records and evidence for a decrease in pCO₂ during the
49
50 621 Oligocene through early Miocene (34–16 Ma). *Palaeogeography, Palaeoclimatology,*
51
52 622 *Palaeoecology* 260, 41-49.
53
54
55
56
57
58
59
60
61
62
63
64
65

- 623 Popova, S., Utescher, T., Gromyko, D., Bruch, A., Mosbrugger, V., 2012. Palaeoclimate
1 evolution in Siberia and the Russian Far East from the Oligocene to Pliocene–
2
3 624
4 evidence from fruit and seed floras. *Turkish Journal of Earth Sciences* 21, 315-334.
5 625
6
7 626 Quan, C., Liu, Y.-S., Utescher, T., 2012. Paleogene temperature gradient, seasonal variation
8 and climate evolution of northeast China. *Palaeogeography, Palaeoclimatology,*
9 627
10 *Palaeoecology* 313–314, 150-161.
11 628
12
13 629 Ramstein, G., Fluteau, F., Besse, J., Joussaume, S., 1997. Effect of orogeny, plate motion and
14 land-sea distribution on Eurasian climate change over the past 30 million years.
15 630
16 *Nature* 386, 788-795.
17 631
18
19 632 Roth-Nebelsick, A., Oehm, C., Grein, M., Utescher, T., Kunzmann, L., Friedrich, J.P.,
20 Konrad, W., 2014. Stomatal density and index data of *Platanus neptuni* leaf fossils
21 633
22 and their evaluation as a CO₂ proxy for the Oligocene. *Review of Palaeobotany and*
23 634
24 *Palynology* 206, 1-9.
25 635
26
27 636 Scher, H.D., Whittaker, J.M., Williams, S.E., Latimer, J.C., Kordesch, W.E., Delaney, M.L.,
28 2015. Onset of Antarctic Circumpolar Current 30 million years ago as Tasmanian
29 637
30 Gateway aligned with westerlies. *Nature* 523, 580.
31 638
32
33 639 Seton, M., Müller, R.D., Zahirovic, S., Gaina, C., Torsvik, T., Shephard, G., Talsma, A.,
34 Gurnis, M., Turner, M., Maus, S., Chandler, M., 2012. Global continental and ocean
35 640
36 basin reconstructions since 200 Ma. *Earth-Science Reviews* 113, 212-270.
37 641
38
39 642 Sluijs, A., Schouten, S., Donders, T.H., Schoon, P.L., Röhl, U., Reichart, G.-J., Sangiorgi, F.,
40 Kim, J.H., Damsté, J.S.S., Brinkhuis, H., 2009. Warm and wet conditions in the
41 643
42 Arctic region during Eocene Thermal Maximum 2. *Nature Geoscience* 2, 777-780.
43 644
44
45 645 Spicer, R.A., 2017. Tibet, the Himalaya, Asian Monsoons and Biodiversity-In what ways are
46 they related? *Plant Diversity* 39, 233-244.
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

- 647 Spicer, R.A., Yang, J., Herman, A.B., Kodrul, T., Maslova, N., Spicer, T.E., Aleksandrova,
1
2 648 G., Jin, J., 2016. Asian Eocene monsoons as revealed by leaf architectural signatures.
3
4 649 Earth and Planetary Science Letters 449, 61-68.
5
6
7 650 Spicer, R., Yang, J., Herman, A., Kodrul, T., Aleksandrova, G., Maslova, N., Spicer, T., Ding,
8
9 651 L., Xu, Q., Shukla, A., 2017. Paleogene monsoons across India and South China:
10
11 Drivers of biotic change. *Gondwana Research* 49, 350-363.
12 652
13
14 653 Steininger, F.F., Wessely, G., 2000. From the Tethyan Ocean to the Paratethys Sea:
15
16 Oligocene to Neogene stratigraphy, paleogeography and paleobiogeography of the
17 654
18 circum-Mediterranean region and the Oligocene to Neogene Basin evolution in
19 655
20 Austria. *Mitteilungen der Österreichischen Geologischen Gesellschaft* 92, 95-116.
21 656
22
23 657 Sun, J.M., Ye, J., Wu, W.Y., Ni, X.J., Bi, S.D., Zhang, Z.Q., Liu, W.M., Meng, J., 2010. Late
24
25 Oligocene–Miocene mid-latitude aridification and wind patterns in the Asian interior.
26 658
27
28 *Geology* 38, 515-518.
29 659
30
31 660 Tanai, T., Uemura, K., 1991. The Oligocene noda flora from the Yuya-wan area of the
32
33 western end of Honshu, Japan. I. *Bulletin of the National Science Museum. Series C*
34 661
35
36 662 17, 57-80.
37
38
39 663 Uhl, D., Klotz, S., Traiser, C., Thiel, C., Utescher, T., Kowalski, E., Dilcher, D.L., 2007.
40
41 664 Cenozoic paleotemperatures and leaf physiognomy—a European perspective.
42
43 *Palaeogeography, Palaeoclimatology, Palaeoecology* 248, 24-31.
44 665
45
46 666 Utescher, T., Bruch, A.A., Erdei, B., François, L., Ivanov, D., Jacques, F.M.B., Kern, A.K.,
47
48 Liu, Y.S., Mosbrugger, V., Spicer, R.A., 2014. The Coexistence Approach—
49 667
50
51 668 Theoretical background and practical considerations of using plant fossils for climate
52
53 669 quantification. *Palaeogeography, Palaeoclimatology, Palaeoecology* 410, 58-73.
54
55
56 670 Utescher, T., Mosbrugger, V., Ashraf, A.R., 2000. Terrestrial Climate Evolution in Northwest
57
58 671 Germany Over the Last 25 Million Years. *PALAIOS* 15, 430-449.
59
60
61
62
63
64
65

- 672 Valdes, P., 2000. Paleoclimate modeling, in: Mote, P., O'Neill, A. (Eds.), Numerical
1
2 673 modeling of the global atmosphere in the climate system. NATO Science Series:
3
4 674 Mathematical and Physical Sciences, pp. 465-488.
5
6
7 675 Valdes, P., Armstrong, E., Badger, M., Bradshaw, C., Bragg, F., Crucifix, M., Davies-
8
9 676 Barnard, T., Day, J., Farnsworth, A., Gordon, C., 2017. The BRIDGE HadCM3
10
11 677 family of climate models: HadCM3@ Bristol v1. 0, Geosci. Model Dev., 10, 3715–
12
13 678 3743.
14
15
16
17 679 Wu, M., Chan, J.C., 1997. Upper-level features associated with winter monsoon surges over
18
19 680 South China. Monthly Weather Review 125, 317-340.
20
21
22 681 Xing, Y.W., Gandolfo, M.A., Onstein, R.E., Cantrill, D.J., Jacobs, B.F., Jordan, G.J., Lee,
23
24 682 D.E., Popova, S., Srivastava, R., Su, T., 2016. Testing the Biases in the Rich
25
26 683 Cenozoic Angiosperm Macrofossil Record. International Journal of Plant Sciences
27
28 684 177, 371-388.
29
30
31
32 685 Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms, and
33
34 686 aberrations in global climate 65 Ma to present. Science 292, 686-693.
35
36
37 687 Zachos, J.C., Dickens, G.R., Zeebe, R.E., 2008. An early Cenozoic perspective on
38
39 688 greenhouse warming and carbon-cycle dynamics. Nature 451, 279-283.
40
41
42 689 Zanazzi, A., Kohn, M.J., MacFadden, B.J., Terry, D.O., 2007. Large temperature drop across
43
44 690 the Eocene–Oligocene transition in central North America. Nature 445, 639-642.
45
46
47 691 Zhang, Y.G., Pagani, M., Liu, Z., Bohaty, S.M., DeConto, R., 2013. A 40-million-year
48
49 692 history of atmospheric CO₂. Phil. Trans. R. Soc. A 371, 20130096.
50
51
52 693 Zhang, Z.S., Wang, H.J., Guo, Z.T., Jiang, D.B., 2007. What triggers the transition of
53
54 694 palaeoenvironmental patterns in China, the Tibetan Plateau uplift or the Paratethys
55
56 695 Sea retreat? Palaeogeography, Palaeoclimatology, Palaeoecology 245, 317-331.
57
58
59 696
60
61
62
63
64
65

697 **Figure Captions:**

698

699 **Fig. 1** Comparison of temperature (median values of MAT, CMT, WMT) patterns between
700 the Early (left hand panels) and Late (right hand panels) Oligocene in Eurasia. Topography
701 was derived from the Robertsons paleogeographic data. The grey color lines and dark thick
702 lines on the map represent the present day and Oligocene land–water boundaries respectively.

703

704 **Fig. 2** Comparison of precipitation (median values of MAT, DryMP, WetMP) patterns
705 between the Early (left hand panels) and the Late (right hand panels) Oligocene in Eurasia.
706 The geographic map was derived from Robertsons paleogeographic data. The grey color lines
707 and dark thick lines on the map represent the present day and Oligocene land–water
708 boundaries respectively.

709

710 **Fig. 3** Climate anomalies of the Oligocene and the present day observations in Eurasia. The
711 anomalies are calculated by Oligocene minus the present day for MAT (A), CMT (B), WMT
712 (C), MAP (D), DryMP (E) and WetMP (F) respectively. The geographic map (31 Ma) was
713 derived from Robertsons paleogeographic data. The grey color lines on the map represent the
714 present land–water boundaries.

715

716 **Fig. 4** The differences between models and all fossil data for the Rupelian period. The Y axes
717 refer to the differences between modelling results and the fossil climate data (mean value of
718 model data minus fossil data) for MAT, CMT, WMT, MAP, DryMP and WetMP respectively.
719 The X axes indicate different simulation names. Different bar colors indicate different CO₂
720 level: green, yellow and red colors represent 560 ppmv, 840 ppmv, and 1120 ppmv
721 respectively. Grey color bars are the mean values of all the models. Different fill pattern

722 represent different geographic models, the blank and diagonal fill bars represent Robertsons
723 and Getech Group Plc. Simulations respectively.

724
725 **Fig. 5** Difference between simulations and the fossil data for the Chattian period. The Y axes
726 refers to the differences between modelling results and the fossil climate data (mean value of
727 model data minus fossil data) for MAT, CMT, WMT, MAP, DryMP and WetMP respectively.
728 The X axes show different simulation names. Different bar colors indicate different CO₂ level:
729 blue, green, yellow and red colors represent 280 ppmv, 560 ppmv, 840 ppmv and 1120 ppmv
730 respectively. Grey color bars represent mean values of all the models except ‘tedjm’ and
731 ‘tecqa’. Different fill pattern bars represent different geographic models, the blank and
732 diagonal fill bars represent Robertsons and Getech Group Plc. Simulations respectively.

733
734 **Fig. 6** The Rupelian (31 Ma) fossil data versus the ‘tecqn1’ HadCM3L simulation for MAT
735 (A), CMT (B), WMT (C), MAP (D), DryMP (E) and WetMP (F). Grey color lines and dark
736 thick lines on the map represent the present day and Oligocene land–sea boundaries
737 respectively. Filled circles represent the fossil data while contoured map indicates the
738 simulation data.

739
740 **Fig. 7** The Chattian (26 Ma) fossil data versus the ‘tedjp’ HadCM3L simulation. for MAT
741 (A), CMT (B), WMT (C), MAP (D), DryMP (E) and WetMP (F). Grey color lines and dark
742 thick lines on the map represent the present day and Oligocene land–sea boundaries
743 respectively. Filled circles represent the fossil data while contoured map indicates the
744 simulation data.

745

746 **Fig. 8** Vector winds and strength (m/s) at 850 hPa during the Rupelian ('tecqn1' simulation)

747 model in JJA (A) and DJF (B).

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

Figure 2
[Click here to download high resolution image](#)

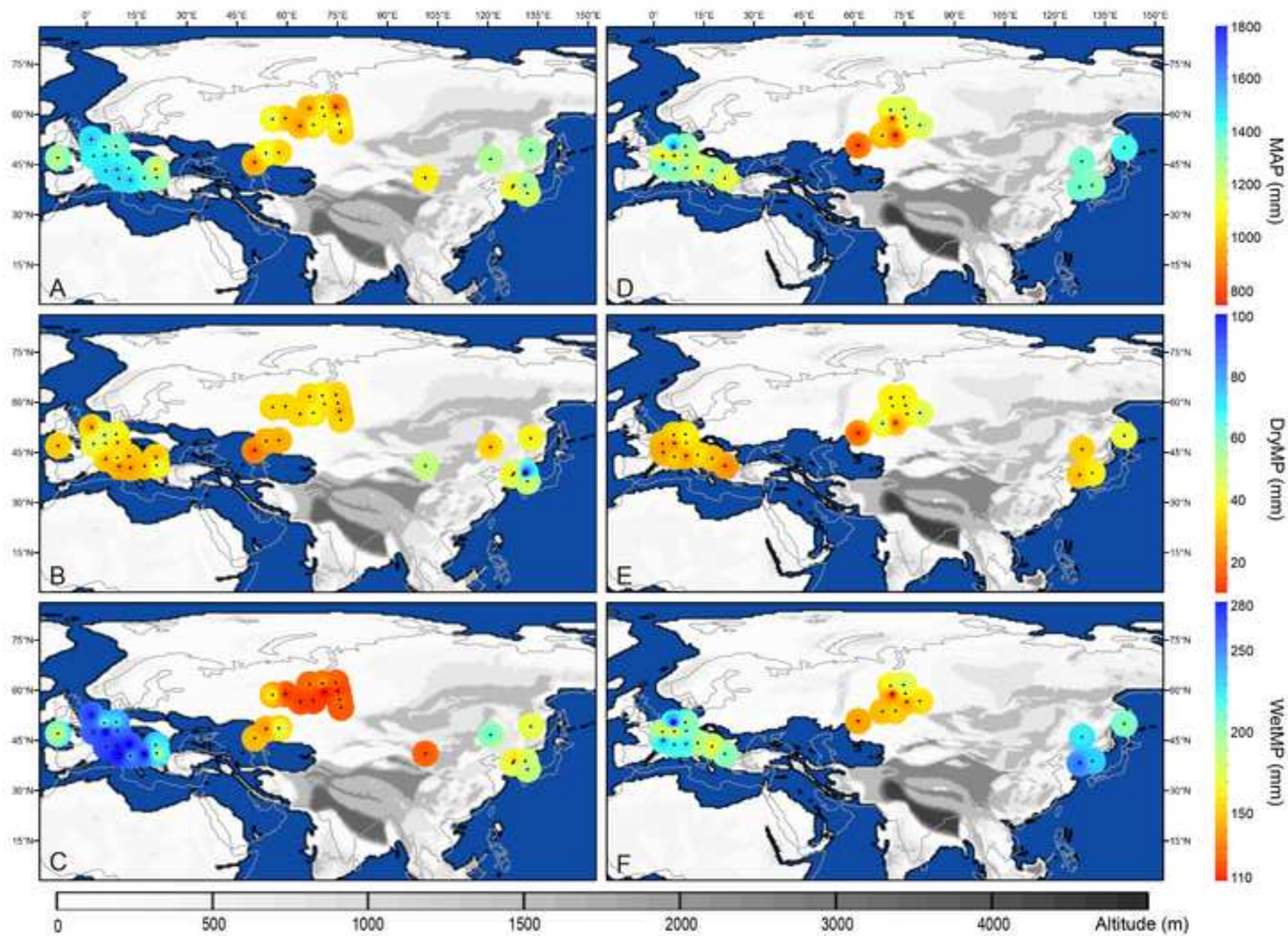


Figure 3
[Click here to download high resolution image](#)

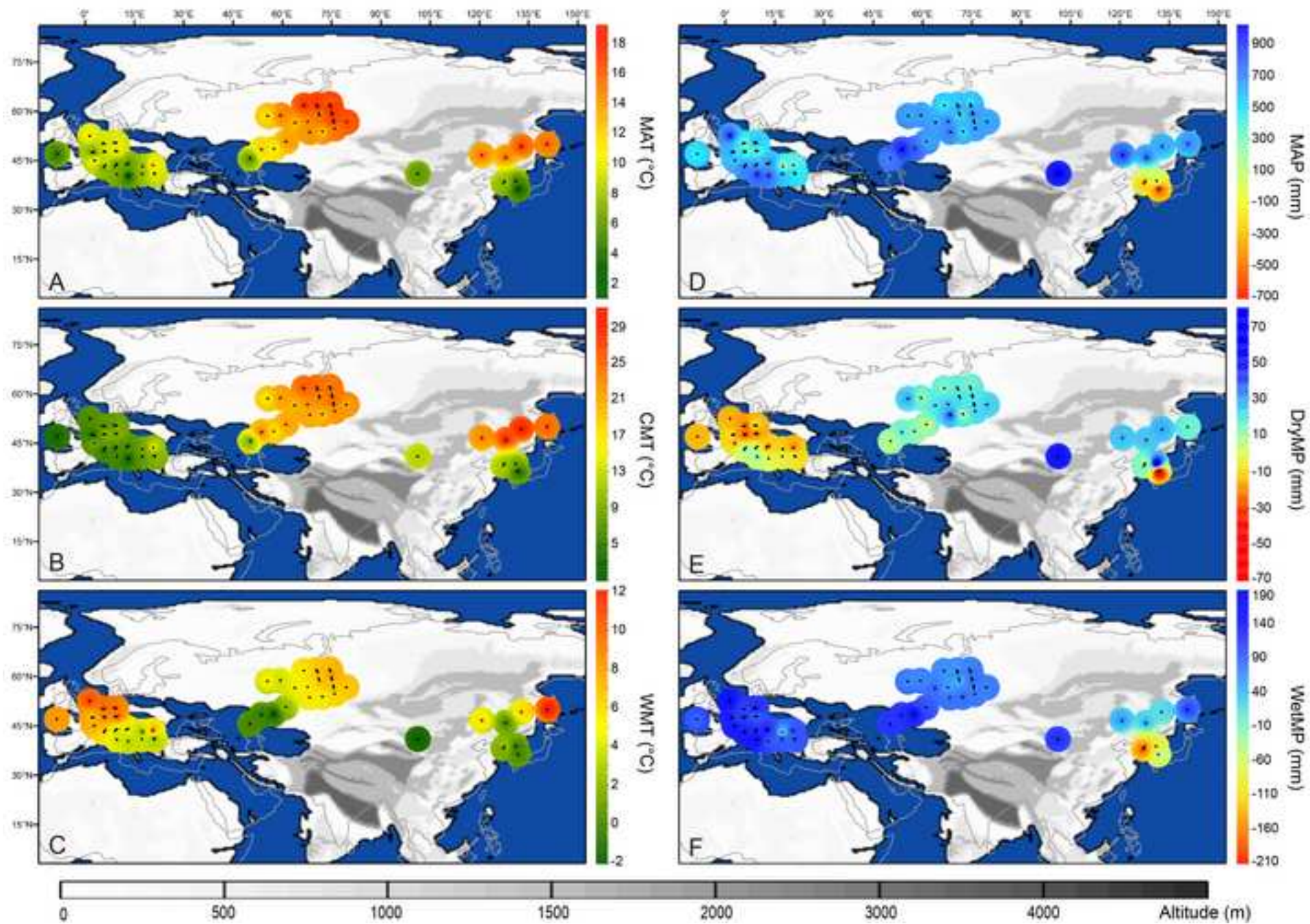


Figure 4

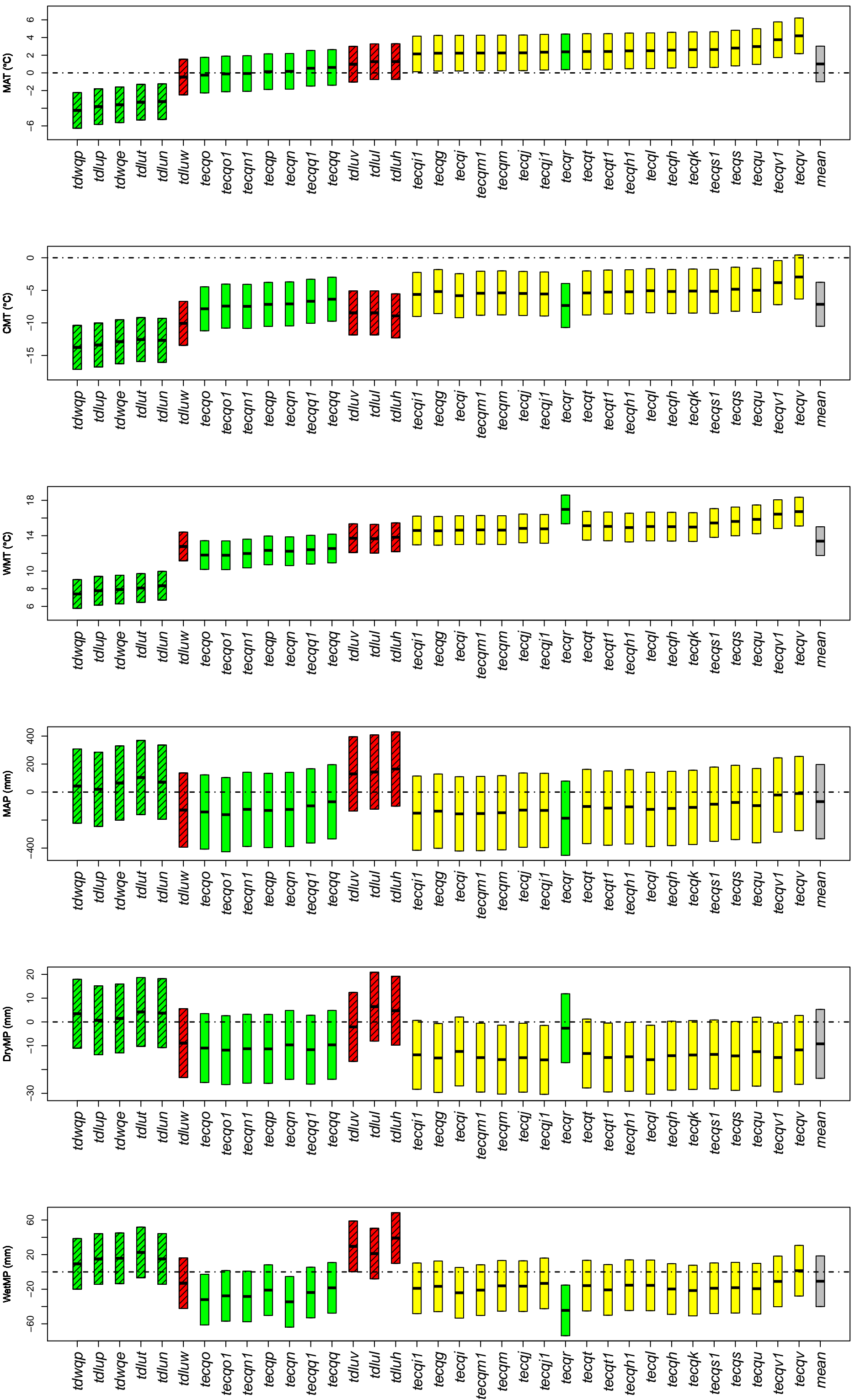


Figure 5

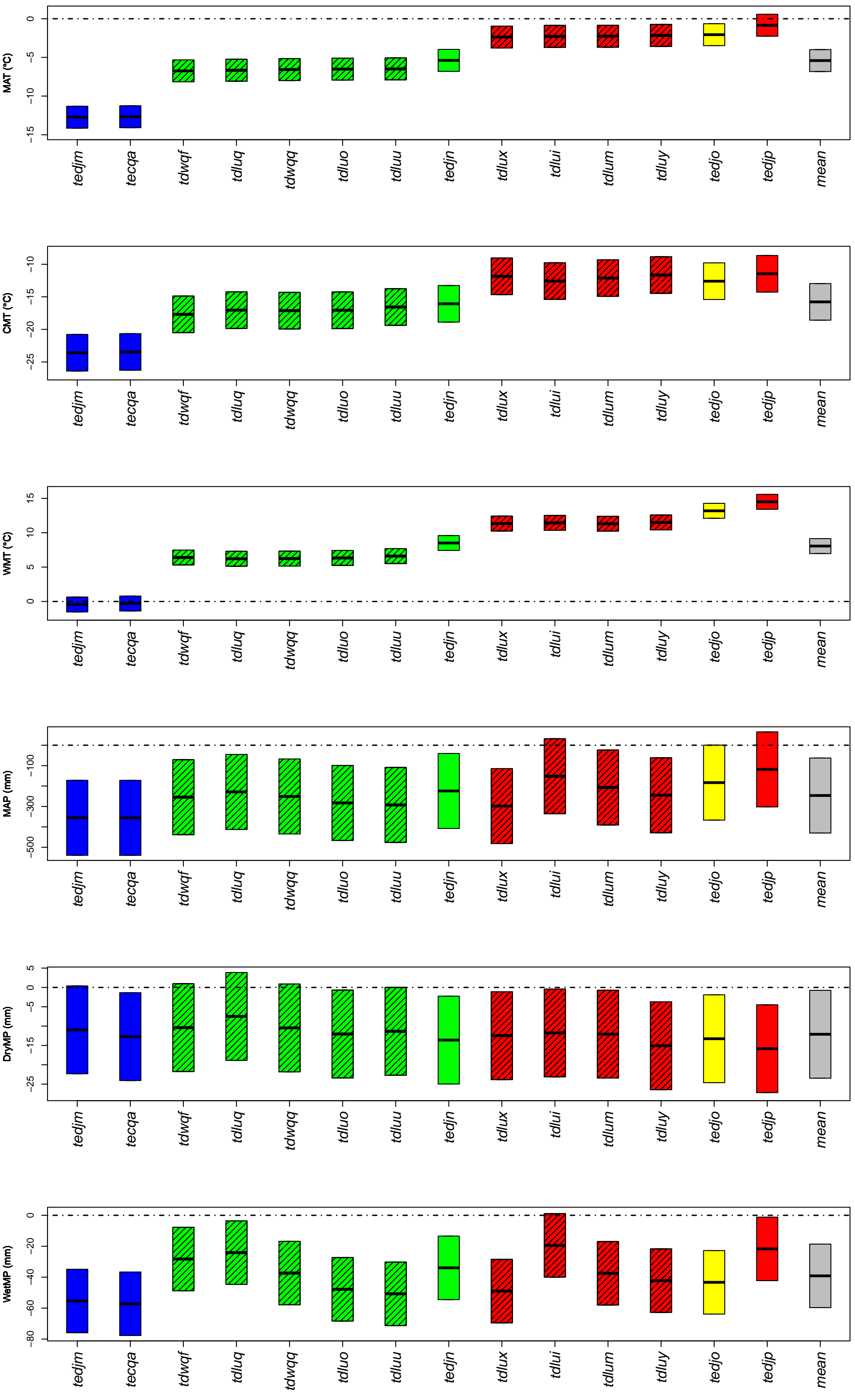


Figure 6
[Click here to download high resolution image](#)

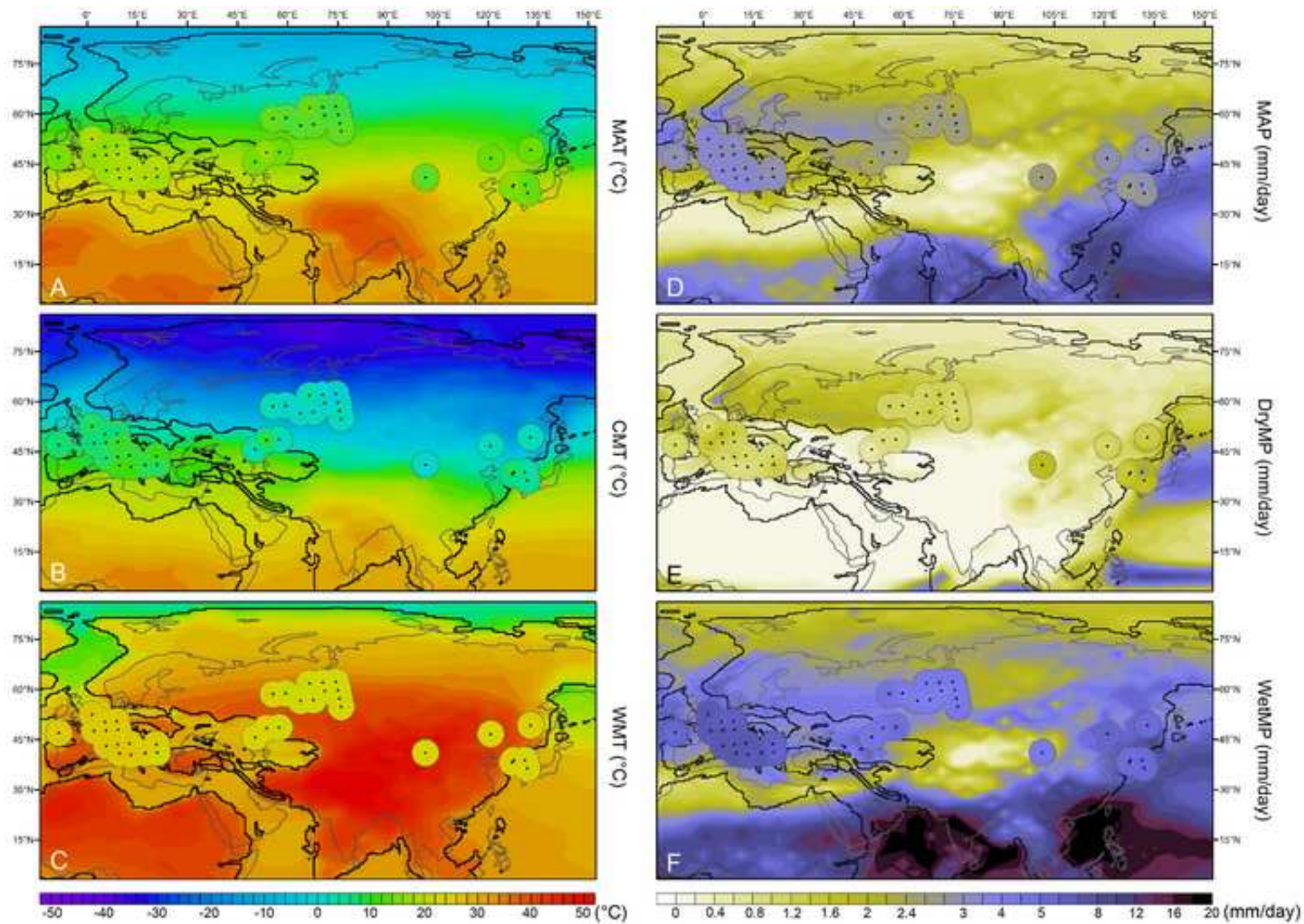


Figure 7
[Click here to download high resolution image](#)

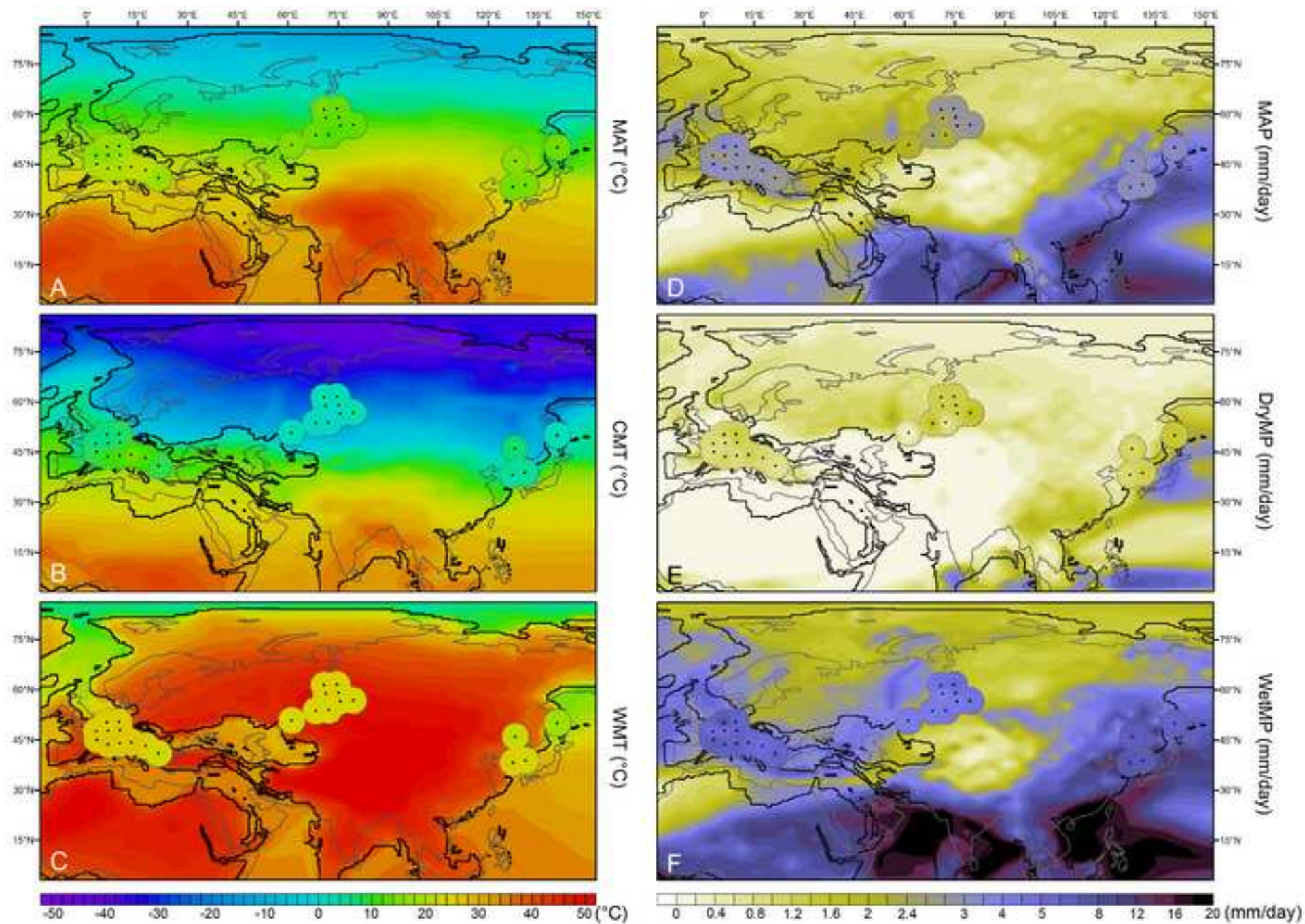


Figure 8
[Click here to download high resolution image](#)

