

*Anthropogenic impacts on recent decadal change in temperature extremes over China: relative roles of greenhouse gases and anthropogenic aerosols*

Article

Accepted Version

Chen, W. and Dong, B. (2019) Anthropogenic impacts on recent decadal change in temperature extremes over China: relative roles of greenhouse gases and anthropogenic aerosols. *Climate Dynamics*, 52 (5-6). ISSN 0930-7575 doi: <https://doi.org/10.1007/s00382-018-4342-9> Available at <http://centaur.reading.ac.uk/78120/>

It is advisable to refer to the publisher's version if you intend to cite from the work. See [Guidance on citing](#).

To link to this article DOI: <http://dx.doi.org/10.1007/s00382-018-4342-9>

Publisher: Springer

All outputs in CentAUR are protected by Intellectual Property Rights law, including copyright law. Copyright and IPR is retained by the creators or other copyright holders. Terms and conditions for use of this material are defined in the [End User Agreement](#).

[www.reading.ac.uk/centaur](http://www.reading.ac.uk/centaur)

## **CentAUR**

Central Archive at the University of Reading

Reading's research outputs online

**Anthropogenic impacts on recent decadal change in temperature extremes over**

**China: relative roles of greenhouse gases and anthropogenic aerosols**

Wei Chen<sup>1</sup> and Buwen Dong<sup>2</sup>

<sup>1</sup> State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100029

<sup>2</sup> National Centre for Atmospheric Science-Climates, Department of Meteorology, University of Reading, Reading, UK

Corresponding author:

Wei Chen,

Institute of Atmospheric Physics,

Chinese Academy of Sciences,

P. O. Box 9804, Beijing 100029, China.

E-mail: [chenwei@mail.iap.ac.cn](mailto:chenwei@mail.iap.ac.cn)

## Abstract

Observational analysis indicates significant changes in some temperature extremes over China across the mid-1990s. The decadal changes in hot extremes are characterized as a rise in annual hottest day and night temperatures (TXx and TNx) and an increase in frequencies of summer days (SU) and tropical night (TR). The decadal changes in cold extremes are distinguished by a rise in annual coldest day and night temperatures (TXn and TNn) and a decrease in frequencies of ice days (ID) and frost days (FD). These decadal changes manifest not only over China as a whole, but also over individual climate sub-regions.

An atmosphere-ocean-mixed-layer coupled model forced by changes in greenhouse gases (GHG) concentrations and anthropogenic aerosol (AA) emissions realistically reproduces the general spatial patterns and magnitudes of observed changes in both hot and cold extremes across the mid-1990s, suggesting a pronounced role of anthropogenic changes in these observed decadal changes. Separately, changes in GHG forcing lead to rise in TXx, TNx, TXn and TNn, increase in frequencies of SU and TR and decrease in frequencies of ID and FD over China through increased Greenhouse Effect with positive clear sky longwave radiation and play a dominant role in simulated changes of both hot and cold extremes over China. The AA forcing changes tend to cool Southern China and warm Northern China during summer via aerosol-radiation interaction and AA-induced atmosphere-cloud feedback and therefore lead to some weak increase in hot extremes over Northern China and decrease over Southeast China.

22 Meanwhile, AA changes lead to warming over China during winter through cloud  
23 feedbacks related to aerosol induced cooling over tropical Indian Ocean and western  
24 tropical Pacific, and also induce changes in cold extremes the same sign as those  
25 induced by GHG, but with weak magnitude.

26 **Key words:** hot temperature extremes; cold temperature extremes; China; decadal  
27 change; the mid-1990s; greenhouse gases; anthropogenic aerosol

## 28 **1. Introduction**

29 Understanding of the changes in climate extremes and the underlying drivers is  
30 important for human society, economies and ecosystems. In the last few decades,  
31 temperature extremes exhibited robust changes at global and regional scales, with more  
32 hot extremes and less cold extremes (e.g., Alexander et al. 2006; Donat et al. 2013).  
33 The impacts of temperature extremes have highlighted the urgency of improved  
34 understanding of their physical causes and to what extent they are manifested in a  
35 warming world (e.g., Otto et al. 2012; Christidis et al. 2013; Perkins 2015).

36 China experienced record-breaking heat waves and temperature extremes that  
37 imposed disastrous impacts on individuals and society (e.g., Yin et al. 2016; Zhou et al.  
38 2016; Freychet et al. 2017). Such as the 2013 heat wave in Central and Eastern China  
39 (Ma et al. 2017), the 2014 hot and dry summer over Northeast China (Wilcox et al.  
40 2015) and the 2015 extreme hot summer over Western China (Sun et al. 2016). The  
41 trend of continuous warming and increase in hot extremes over China might be

42 associated with the global-scale warming (e.g., Wei et al. 2011). This warming trend  
43 and increase in hot temperature extremes can be reproduced by the future climate  
44 change scenario (Yao et al. 2012), implying the role of anthropogenic activity in  
45 increasing hot temperature extremes.

46 Previous attribution studies detected that anthropogenic activity, as a combined  
47 effects of greenhouse gases (GHG) concentrations and anthropogenic aerosol (AA)  
48 emissions, induces the changes in temperature extremes over China. Approximately 90%  
49 of the observed changes in hot extremes since mid-20th century may contributed by  
50 anthropogenic forcing (e.g., Wen et al. 2013; Yin et al. 2016). The summer mean  
51 temperature and temperature extremes in Eastern China can be increased by the  
52 anthropogenic influence (Sun et al. 2014). The direct impacts of changes in GHG  
53 concentrations and AA contribute to the 2014 hot and dry summer in Northeast China,  
54 beside SST anomaly (Wilcox et al. 2015). Both anthropogenic factors and atmospheric  
55 natural variability contributed to the 2013 mid-summer heat wave in Central and  
56 Eastern China (Ma et al. 2017).

57 Physically, the climate system warms in response to the increase in GHG  
58 concentrations, because the atmosphere traps more outgoing longwave radiation (e.g.,  
59 Cubasch et al. 2001; Dong et al. 2009, 2017a). In addition, AA can affect the surface  
60 and atmospheric temperature by altering the radiative properties of clouds through  
61 aerosol-cloud interaction (Hansen et al. 1997; Stevens and Feingold 2009), and by  
62 scattering and absorbing the solar radiation directly through aerosol-radiation

63 interaction. Additionally, remote AA emissions can impact on local temperature and  
64 temperature extremes through changing dynamics. For example, the remote AA  
65 emissions over Europe have a downstream extension impact on the temperature and  
66 temperature extremes over East Asia (Dong et al. 2015, 2016). Besides anthropogenic  
67 aerosol emissions, natural aerosol emissions can also influence climate dynamics (e.g.,  
68 Yang et al., 2016, 2017).

69 The previous studies have highlighted external forcings, particularly anthropogenic  
70 changes, play an important role in decadal changes of temperature extremes. However,  
71 the relative individual contributions of changes in GHG concentrations and AA  
72 emissions to the observed changes in temperature extremes are still not clear. Therefore,  
73 the main aims of this work are to quantify the relative roles of changes in GHG and AA  
74 forcing in shaping the changes in temperature extremes over China, and to understand  
75 the physical processes responsible.

76 Despite the rapid development of attribution studies in climate extremes in recent  
77 years (Stott et al. 2016), there is still no consensus about the best methodology for  
78 attribution. One widely-used attribution approach relies on an atmospheric general  
79 circulation model (AGCM) forced by prescribed sea surface temperatures (SSTs), with  
80 and without anthropogenic influences (e.g., Christidis et al. 2013; Kamae et al. 2014;  
81 Kim et al. 2015; Schaller et al. 2016). A potential limitation of these experiments is the  
82 lack of explicit air–sea interaction, which causes an inconsistency in surface energy  
83 fluxes and can limit a model’s ability to accurately simulate natural climate variability

84 (e.g., Barsugli and Battisti 1998; He and Soden 2016). Another ordinary attribution  
85 method is based on a coupled general circulation model (CGCM) with constant  
86 emissions, which reaches equilibrium after a long integration (Bollasina et al. 2011; Li  
87 and Ting 2016; Wang et al. 2012, 2013). The experiments in CGCMs with full ocean  
88 dynamics have huge computational cost. Moreover, the CGCMs may exhibit significant  
89 biases in the mean state, such as a large cold equatorial SST bias in Pacific (Vanniere  
90 et al. 2012). Thus, replacing the three-dimension ocean GCM with an ocean mixed-  
91 layer model would reduce the cost of the experiments, and have a smaller SST bias (due  
92 to a prescribed flux correction), whilst also retaining intra-seasonal variability and  
93 coupling between the atmosphere and the ocean. Therefore, this work is based on a set  
94 of experiments using an atmosphere-ocean-mixed-layer couple model (Hirons et al.  
95 2015; Tian et al. 2018).

96 The structure of this paper is organized as follows: Section 2 illustrates the  
97 observed decadal changes in temperature extremes over China. The model and  
98 experiments are described briefly in Section 3. Section 4 evaluates the simulated  
99 changes in response to changes in GHG concentrations and AA emissions. Sections 5  
100 and 6 illustrate the physical processes involved in the responses of hot and cold  
101 extremes to changes in anthropogenic forcings, respectively. Conclusions are  
102 summarized in Section 7.



## 103 **2. Observed decadal changes in temperature extremes over China**

### 104 **2.1 Observational datasets**

105 The China stations data used are the homogenized datasets of daily maximum  
106 temperature (Tmax) and minimum temperature (Tmin) series with 753 stations in China  
107 from 1960 to 2016 (Li et al. 2016). Considering various climate types in China, we  
108 divide the 753 stations into three sub-regions: Northern China (NC) with 331 stations  
109 north of 35°N, Southeastern China (SEC) with 334 stations south of 35°N and east of  
110 105°E, and Southwestern China (SWC) with 88 stations south of 35°N and west of  
111 105°E. The distributions of these three sub-regions are shown in Fig. 2a. Also used are  
112 the global land gridded climate extremes (GHCNDEX) based on the Global Historical  
113 Climatology Network (GHCN)-Daily dataset from 1960 to 2011 (Donat et al. 2013).  
114 The hot extremes indices are annual hottest day temperature (TXx), and warmest night  
115 temperature (TNx), the frequency of summer days (SU, annual number of days when  
116 Tmax >25°C), and tropical night (TR, annual number of days when Tmin >20 °C). The  
117 cold extremes indices are annual coldest day temperature (TXn), and coldest night  
118 temperature (TNn), the frequency of ice days (ID, annual number of days when Tmax  
119 <0 °C), and frost days (FD, annual number of days when Tmin <0 °C).

### 120 **2.2 Observed decadal changes since the mid-1990s**

121 Figure 1 illustrates the time evolution of the area averaged annual mean  
122 temperature extremes anomalies over China and over three sub-regions, relative to the

123 climatology, averaged over the whole time period. These time series clearly show the  
124 abrupt changes in both hot and cold extremes since the mid-1990s. Therefore, the  
125 decadal change in this study is compared between present day (PD) of 1994~2011 and  
126 early period (EP) of 1964~1981. During summer, a rapid increase in TXx since the mid-  
127 1990s occurs in China (Fig. 1a). The change in TXx anomaly during the PD relative to  
128 the EP is 0.57 °C in China station data, which is about two times as large as its  
129 interannual variation of 0.27 °C. This robust increase is also supported by the  
130 GHCNDEX data with a change in TXx anomaly of 0.77 °C. Additionally, the increase  
131 of TXx from the EP to the PD occurs in each sub-region over China, with a range of  
132 changes from 0.40 °C to 0.76 °C. Moreover, accompanied with the increase of TXx, the  
133 frequency of SU rises by 9 days over China (8 days in GHCNDEX data; Fig. 1c).  
134 Similar to the increase of TXx, TNx also shows significant increase over China since  
135 the mid-1990s (Fig. 1b). The change in TNx anomaly is 0.76 °C (0.96 °C in GHCNDEX  
136 data). The remarkable decadal increase of TNx occurs in individual sub-regions over  
137 China, with the greatest amplitude of 1.05 °C over NC and the smallest amplitude of  
138 0.52 °C over SEC. The frequency of TR rises by 8 days in China (Fig. 1d), coinciding  
139 with the increase of TNx.

140 During winter, the cold extremes also exhibit decadal changes since the mid-1990s,  
141 being characterized as a rise in temperature and a decrease in frequency of cold days.  
142 TXn anomaly increases by 1.48 °C over China (Fig. 1e), being similar to the change of  
143 1.45 °C in GHCNDEX data. The increase of TXn manifests over three sub-regions with

144 a range from 0.94 °C to 1.59 °C. As a result of the increase of TX<sub>n</sub>, the frequency of ID  
145 is decreased by about 4 days over China (Fig. 1g). The decrease in ID is mainly over  
146 NC, with the magnitude of 8 days. Moreover, a robust increase in TN<sub>n</sub> appears since  
147 the mid-1990s. The changes of TN<sub>n</sub> anomaly is 1.82 °C over China, being consistent  
148 with 1.86 °C in GHCNDEX data (Fig. 1f). The greatest increase of TN<sub>n</sub> is over NC  
149 (2.17 °C). Additionally, the frequency of FD is decreased by about 10 days (Fig. 1h). It  
150 is noted that the changes in cold extremes are larger than hot extremes. This is consistent  
151 with stronger seasonal warming over northern hemisphere mid-latitudes in boreal  
152 winter than in boreal summer in response to anthropogenic forcing (e. g., John et al.  
153 2012; Dong et al. 2017b), which is related to the snow-albedo feedback (e.g., Stouffer  
154 and Wetherald 2007; Rangwala et al. 2016).

155 The spatial patterns of changes in these temperature extremes during the PD  
156 relative to the EP are illustrated in Fig. 2. The most important features of changes are  
157 the increase in hot extremes and decrease in cold extremes over the most regions of  
158 China although there are some spatial variations (Figs. 2b-i). For the hot extremes, the  
159 changes in TX<sub>x</sub> and TN<sub>x</sub> show a large increase over NC with a magnitude of about  
160 1.0~1.5 °C (Figs. 2b and c). While the changes in the frequencies of SU and TR show  
161 the increase in a large domain over SEC (Figs. 2d and e). For the cold extremes, the  
162 TX<sub>n</sub> and TN<sub>n</sub> show a relatively uniform increase over China with a range of 1.5~2.5  
163 °C (Figs. 2f and g). The frequencies of ID and FD show a decrease over the most regions  
164 of China (Figs. 2h and i).

165 The robust decadal changes in the temperature extremes have been observed over  
166 China since the mid-1990s. Questions come out naturally: what has caused these rapid  
167 changes? Do the anthropogenic activities drive these changes? A set of experiments  
168 using a coupled climate model are performed to assess contributions of changes in  
169 anthropogenic forcings (GHG concentrations and AA emissions) to observed decadal  
170 changes in temperature extremes over China since the mid-1990s, and to quantify the  
171 relatively roles of individual forcing factors and to elucidate physical processes  
172 involved.

### 173 **3. Model and experiments design**

174 The model used is an atmosphere-ocean-mixed-layer coupled model called  
175 MetUM-GOML1 (Hirons et al. 2015). The atmospheric component is the Met Office  
176 Unified Model (MetUM) at the fixed scientific configuration Global Atmosphere 3.0  
177 (GA3.0; Arribas et al. 2011; Walters et al. 2011) with a horizontal resolution of  $1.875^\circ$   
178 longitude and  $1.25^\circ$  latitude.

179 The model includes earth system components such as an interactive tropospheric  
180 aerosol scheme and the following aerosol: ammonium sulphate, mineral dust, sea salt,  
181 fossil fuel black carbon, fossil fuel organic carbon, biomass burning aerosols, and  
182 secondary organic (biogenic) aerosols. The direct radiative effect due to scattering and  
183 absorption of radiation by all aerosol species is represented in the model. The semi-  
184 direct effect, whereby aerosol absorption tends to change cloud formation by warming  
185 the aerosol layer, is included implicitly (Walters et al. 2011). The parameterization of

186 the indirect effects is described in detail by Jones et al. (2011). The model validation  
187 suggested a good performance in simulating aerosol properties and the detailed  
188 description of this aspect has been documented in Bellouin et al. (2011). The modeled  
189 sulphate aerosol surface concentrations, nitrate aerosol concentrations, carbonaceous  
190 aerosol concentrations and total AODs are all compares well against observed  
191 measurements. Moreover, the model reproduced the known pattern of AOD, with  
192 industrial pollution in North America, Europe, and Asia, biomass burning aerosols in  
193 Central Africa and South America, and mineral dust transport across the Atlantic and  
194 Arabian Sea. The eastward gradient in AOD in North America and China is well  
195 reproduced.

196 The oceanic component is a Multi-Column K Profile Parameterization (MC-KPP)  
197 mixed-layer ocean model. The atmospheric and oceanic components are coupled every  
198 three hours. The air–sea coupling is limited by the maximum extent of a seasonally  
199 varying sea ice climatology (Hirons et al. 2015). In the uncoupled region of MetUM-  
200 GOML1, the atmosphere is forced by the repeating mean annual cycle of SST and sea  
201 ice extent (SIE) from the Met Office HadISST data set (Rayner et al. 2003). The  
202 horizontal resolution of MC-KPP is the same as the MetUM where it is coupled. The  
203 MC-KPP columns have 100 vertical levels with a depth of 1000m. The vertical  
204 discretization allows very high resolution (approximately one meter) in the upper ocean.  
205 Since MC-KPP simulates only vertical mixing and does not include ocean dynamics,  
206 climatological seasonal cycles of depth-varying temperature and salinity corrections are

207 prescribed to represent the mean ocean advection and account for biases in atmospheric  
208 surface fluxes.

209 Since the mid-1990s, there have been increased anthropogenic GHG  
210 concentrations (14% increase in CO<sub>2</sub>, 23% increase in CH<sub>4</sub> and 7% increase in N<sub>2</sub>O),  
211 and significant changes in AA emissions. The changes in annual mean sulfur dioxide  
212 emissions are characterized as decreases over Europe and North America and increases  
213 over East and South Asia (Fig. 3).

214 As summarized in Table 1, a 12 year MetUM-GOML1 relaxation experiment (R0)  
215 was firstly performed in which the MC-KPP profiles of temperature and salinity were  
216 relaxed to a present day (PD; 1994~2011) ocean temperature and salinity climatology  
217 derived from the Met Office ocean analysis (Smith and Murphy 2007). The relaxation  
218 experiment used PD GHG and AA forcings (Lamarque et al. 2010, 2011). The daily  
219 mean seasonal cycle of ocean temperature and salinity corrections from the coupled  
220 relaxation experiment are then imposed in free-running coupled experiments. Four  
221 other experiments are performed by using different forcings. These experiments  
222 represent the early period (EP; 1964~1981), All Forcing present day (PDGA), GHG  
223 forcing (PDG) and AA forcing (PDA) with no relaxation. All experiments are run for  
224 50 years and use the climatological PD sea ice extent from HadISST (Rayner et al.  
225 2003). The last 45 years of each experiment are used for analysis. Using the same set  
226 of experiments, Tian et al. (2018) has investigated the responses of the East Asian  
227 summer monsoon.

228 The response to a particular forcing is estimated by the difference between a pair  
229 of experiments that include and exclude that forcing. The combined effect of changes  
230 in both GHG and AA (hereafter All forcing) is the difference between PDGA and EP  
231 experiment (PDGA - EP). The impact of changes in GHG concentrations (hereafter  
232 GHG forcing) is the difference between PDG and EP (PDG - EP) and the impact of  
233 changes in AA emissions (hereafter AA forcing) is the difference between PDA and EP  
234 (PDA - EP).

#### 235 **4. Model simulated changes in response to anthropogenic forcing**

236 The spatial distributions of changes in hot extremes in response to different  
237 forcings are shown in Fig. 4. The model experiment in response to changes in All  
238 forcing from the EP to the PD, not only reproduces the significant increase of hot  
239 extremes over China, but also captures the generally spatial patterns of observed  
240 changes (Figs. 4a-d). The increase of TXx and TNx in response to All forcing changes  
241 exceeds 0.5 °C over the most area of China (Figs. 4a and b), with a maximum center  
242 over NC (exceed 1.0 °C), being consistent with observations (Figs. 2b and c). As a result,  
243 the frequencies of SU and TR manifest a significant increase over China in response to  
244 All forcing changes (Figs. 4c and d). The large increase domain is over SEC, which is  
245 also seen in observations (Figs. 2d and e). The similarities between the changes in  
246 response to All forcing and observed changes indicate that the observed increase in hot  
247 extremes over China since the mid-1990s is predominantly due to the anthropogenic

248 GHG and AA changes.

249 Moreover, in response to the GHG forcing change, the hot extremes show a more  
250 or less uniform increase over China (Figs. 4e-h). The spatial pattern and magnitude of  
251 changes in hot extremes in response to GHG forcing changes are similar to those in  
252 response to All forcing changes, indicating that changes in GHG concentrations play a  
253 dominant role in the increase in hot extremes over China. Nevertheless, the role of  
254 changes in AA forcing in the hot extremes is relatively weak and shows a dipole pattern  
255 with increases in north and decreases in south (Figs. 4i-l). The increase in TXx and TNx,  
256 as well as the increase in the frequencies of SU and TR, is shown over NC in response  
257 to changes in AA forcing, although the magnitude is weaker than that in response to  
258 changes in GHG forcing. However, the decrease in TXx, TNx and frequencies of SU  
259 and TR appears over SEC and SWC in response to changes in AA forcing.

260 In terms of the cold extremes, their responses to changes in different forcings are  
261 illustrated in Fig. 5. The rise in TXn and TNn and the decrease in the frequencies of ID  
262 and FD in response to All forcing changes coincide with observations (Figs. 5a-d). In  
263 response to changes in GHG forcing, the increase in TXn and TNn and the decrease in  
264 the frequencies of ID and FD are not only comparable to those in response to All forcing  
265 changes, but also consistent with those in observations (Figs. 5e-h), suggesting that  
266 GHG forcing changes play a vital role in the observed decadal changes of cold extremes.  
267 Additionally, AA forcing changes also contribute to changes in cold extreme (Figs. 5i-  
268 l), particularly to the rise in TXn and TNn and decrease in frequencies of ID and FD



269 over NC and SEC, although the magnitudes of changes are weaker than those in  
270 response to GHG forcing changes.

271 Quantitatively, the model simulated changes in response to All forcing changes  
272 reproduce the observed changes in temperature extremes over China realistically,  
273 although some extreme indices are overestimated a little bit. In response to All forcing  
274 changes, the area averaged TXx (TNx) over China is 1.05 °C (0.92 °C), which are  
275 comparable to the observed changes of 0.58 °C (0.76 °C). The TXn and TNn averaged  
276 over China in response to All forcing changes are 1.69 °C and 1.45 °C, which are very  
277 close to observed changes of 1.48 °C and 1.82 °C.

278 Moreover, the agreement of model simulated magnitude of changes in extreme  
279 indices with those in observations is not only over China as a whole, but also over  
280 individual sub-regions. Figure 6 gives some area averaged changes in temperature  
281 extreme indices over the three sub-regions for both observations and model simulated  
282 responses. The area averaged changes in TXx in response to All forcing changes are  
283 comparable to observations, although they are overestimated a little bit over NC and  
284 SEC. The simulated TNx changes are also in good agreement with the observations,  
285 particularly over NC. The change of TNx over NC in response to All forcing changes  
286 is 1.07 °C, compared to the observed change of 1.05 °C. Additionally, the increases in  
287 frequencies of SU and TR in response to All forcing changes are very similar to those  
288 in observations in each sub-region. On the other side, the model simulated changes of  
289 cold extremes in response to All forcing changes over the three sub-regions are in good

290 agreement with observations. The TXn (TNn) is 2.26 °C (1.84 °C) over NC, 1.28 °C  
291 (1.07 °C) over SEC and 0.64 °C (0.82 °C) over SWC, which are very close to observed  
292 changes of 1.59 °C (2.17 °C) over NC, 1.52 °C (1.59 °C) over SEC and 0.94 °C (1.36  
293 °C) over SWC. The decrease of frequencies of ID and FD also coincides with  
294 observations.

295 Separately, the model simulated response to GHG forcing changes exhibit increase  
296 in hot extremes over all the three sub-regions. The magnitude of this increase in  
297 response to GHG forcing changes is almost equal to that in response to All forcing  
298 changes, indicating a dominant contribution of changes in GHG concentrations to the  
299 simulated increase in hot extremes over China. On the contrary, the simulated responses  
300 of hot extremes in response to AA forcing changes show positive value over NC but  
301 negative value over SEC, which are consistent with the dipole pattern with increases in  
302 north and decreases in south (Figs. 4i-l). In terms of cold extremes, the model simulated  
303 increase in cold temperatures and decrease in frequencies of cold days result from the  
304 combined effects of changes in both GHG concentrations and AA forcing . The  
305 simulated change of TXn and TNn in response to GHG forcing changes is about two to  
306 three times that in response to AA forcing changes. The simulated changes in TXn  
307 (TNn) are 2.03 °C (1.08 °C) over NC and 1.21 °C (1.71 °C) over SEC in response to  
308 GHG forcing changes, in comparison to 0.76 °C (0.55 °C) over NC and 0.64 °C (0.5 °C)  
309 over SEC in response to AA forcing changes.

310 There is a nonlinearity for some extreme temperature changes in response to

311 changes in GHG concentrations and AA forcing in model simulations, i.e., the sum of  
312 responses to separate forcings is not equal to the response to changes in All forcings  
313 together. This nonlinearity is weak over NC. The increase of extremes over NC in  
314 response to changes in GHG forcing explains up to 75% of the TN<sub>x</sub>, 90% of the TX<sub>n</sub>  
315 and 60% of the TN<sub>n</sub> increase in response to All forcing (assuming linearity). But the  
316 nonlinearity is clearly shown for some temperature extremes over SEC (Fig. 6c), and ID  
317 and FD over SWC (Fig. 6f). The nonlinearity of responses to different forcings has  
318 noticed by previous studies (Feichter et al. 2004; Ming and Ramaswamy 2009;  
319 Shiogama et al. 2012). They suggested that the nonlinear cloud response is likely the  
320 source for this nonlinearity. The response of cloud water content and cloud radiative  
321 effect have strong dependency in the combined forcing experiment than in either of the  
322 individual forcing experiments. In our study, the large nonlinearity is over SEC, where  
323 the water vapor content is high. The high humidity tends to increase the nonlinear cloud  
324 response to the anthropogenic forcing. However, detailed discussion of this nonlinearity  
325 is beyond the scope of this study.

326 The simulated response to changes in All forcing indicates that anthropogenic  
327 changes play an important role in generating observed decadal changes in temperature  
328 extremes. However, the responses to changes in GHG and AA show some different  
329 characteristics. Model results indicate that GHG forcing changes tend to increase both  
330 hot and cold extreme temperatures TX<sub>x</sub>, TN<sub>x</sub>, TX<sub>n</sub>, and TN<sub>n</sub>, increase frequency in  
331 SU and TR, and decrease in ID and FD over China, while AA forcing changes are likely

332 to warm NC and cool SEC during summer and induce surface warming over NC and  
333 SEC during winter. The physical processes responsible for the changes in hot and cold  
334 extremes in response to different forcings are discussed in next sections, respectively.

## 335 **5. Physical processes responsible for the decadal changes of hot extremes**

### 336 **5.1 Induces by GHG forcing changes**

337 The spatial patterns of summer mean changes for the key components of surface  
338 energy balance and related variables induced by changes in GHG forcing are illustrated  
339 in Fig. 7. The direct impact of the increase in GHG concentrations leads to an increase  
340 of clear sky downward longwave (LW) radiation of  $0.94 \text{ W m}^{-2}$  over China (Fig. 7a; as  
341 expected for an increase in the Greenhouse Effect), although part of this increase would  
342 be compensated by increase of upward surface LW radiation ( $-0.49 \text{ W m}^{-2}$ ) since  
343 surface warming (Fig. 7b). The net LW anomaly tends to reflect a balance between the  
344 increase emission from the warmer surface (Fig. 7a) and the negative LW cloud  
345 radiative effect (LW CRE; not shown), as a consequence of reduction in cloud cover  
346 (Figs. 7c and d). The decrease in cloud cover over land is related to the decrease in  
347 relative humidity (not shown) since specific humidity over land increases less  
348 (reduction of evapotranspiration related to the  $\text{CO}_2$  physiological effect and constrained  
349 by ocean warming) than specific humidity at saturation which increases with the  
350 continental surface temperature following the Clausius-Clapeyron relationship (e.g.,  
351 Dong et al. 2009; Boé and Terray 2014). The reduction of cloud cover and decrease of

352 relative humidity, being likely due to the surface warming, lead to positive shortwave  
353 cloud radiative effect (SW CRE; with the value of  $1.62 \text{ W m}^{-2}$ ; Fig. 7e) and positive net  
354 surface shortwave (SW) radiation (with the value of  $0.95 \text{ W m}^{-2}$ ; Fig. 7f) over the most  
355 part of China, which in turn have a positive feedback on surface warming. In summary,  
356 it is the increased Greenhouse Effect that induces the increase in hot extremes over  
357 China, with increase of the net downward clear sky LW radiation, in response to GHG  
358 forcing changes. Moreover, the increase of net surface SW radiation related to positive  
359 SW CRE, being associated with the decrease of cloud cover, has a positive feedback  
360 with the surface warming due to the increase of GHG concentrations, which also  
361 contributes to the increase of hot extremes over China.

## 362 **5.2 Induced by AA forcing changes**

363 The spatial distributions of the summer mean changes for the key components of  
364 surface energy balance and related variables induced by changes in AA emissions are  
365 illustrated in Fig. 8. Changes in aerosol optical depth (AOD) indicate a decrease over  
366 Europe and an increase over East Asia and South Asia (Fig. 8a). Local increase of AOD  
367 over East Asia leads to decrease of net clear sky SW radiation ( $-3.36 \text{ W m}^{-2}$ ) over China  
368 through aerosol-radiation interactions (Fig. 8b). However, the SW CRE changes show  
369 positive anomaly, particularly over NC with a magnitude of  $1.22 \text{ W m}^{-2}$  (Fig. 8c). This  
370 positive SW CRE warms the surface and leads to increase in hot extremes over NC  
371 while the decrease of net surface SW radiation through AA induced net clear sky SW  
372 radiation change cools the surface and leads to decrease in hot extremes over SEC. The

373 positive SW CRE over NC is induced by the decrease of cloud cover (Fig. 8d), which  
374 is related to the decrease of soil moisture (Fig. 8d) and water vapor in the atmosphere  
375 (Fig. 8f). This is consistent with Tian et al. (2018), who suggested a drying over NC  
376 related to a weakening of East Asian summer monsoon (EASM) in response to AA  
377 forcing changes. The weakening of EASM is associated with weaker moisture transport  
378 convergence and reduced precipitation (not shown), soil moisture (Fig. 8e) and  
379 evaporation (not shown) over NC, leading to decrease in water vapor in the atmosphere  
380 (Fig. 8f), which in turn gives rise to the positive SW CRE, as a consequence of decrease  
381 in cloud cover. In summary, direct impact of changes in AA forcing induces a decrease  
382 in clear sky SW radiation, which results in surface cooling over SEC and SWC.  
383 However, the positive SW CRE and reduced upward latent heat flux (not shown),  
384 induced by decrease of cloud cover related to reduction of precipitation over NC and  
385 decrease of soil moisture, tend to warm the surface and contribute to increase in hot  
386 extremes over NC.

## 387 **6. Physical processes responsible for the decadal changes of cold extremes**

### 388 **6.1 Induces by GHG forcing changes**

389 Figure 9 is the spatial distributions of the winter mean changes for the key  
390 components of surface energy balance and related variables induced by changes in  
391 GHG concentrations. The downward clear sky LW radiation is increased over southern  
392 part of China with the value of  $0.66 \text{ W m}^{-2}$  (Fig. 9a), although part of this increase is

393 likely to be compensated by the increase of upward surface LW radiation due to  
394 surface warming (Fig. 9b). The positive change of net clear sky LW radiation is partly  
395 due to direct impact of increase in GHG concentrations and partly due to increases of  
396 water vapor in the atmosphere related to GHG induced ocean warming (Fig. 9c).  
397 Furthermore, the net surface SW radiation is increased over northern part of China (Fig.  
398 9d), which is mainly due to the increase of net clear sky SW radiation (Fig. 9e). The  
399 positive clear sky SW radiation, with a value of  $4.5 \text{ W m}^{-2}$  over NC, results from snow  
400 albedo feedback through the reduction of snow cover and depth due to skin temperature  
401 warming (Fig. 9f; e.g., Robock 1983; Yang et al. 2001; Bony et al. 2006; Qu and Hall  
402 2007; Thackeray and Fletcher 2016). In summary, the positive change in net clear sky  
403 LW radiation due to the Greenhouse Effect and associated water vapor feedback  
404 contribute to the warming over SWC and SEC and leads to changes in cold extremes.  
405 The increase of net surface SW radiation, mainly due to the increase of clear sky SW  
406 radiation related to decrease in snow cover and depth, leads to large changes in day-  
407 time extremes of TXn and ID than night-time extremes of TNn and FD over NC.

## 408 **6.2 Induced by AA forcing changes**

409 Figure 10 is the spatial distributions of the winter mean changes for the key  
410 components of surface energy balance and related variables induced by changes in AA  
411 emissions. During winter, AA are advected by mean flow to the Indian Ocean and  
412 western North Pacific, instead of that the AA effects are more over emission area due  
413 to relatively weak advections during summer. There is significant cooling over the

414 Indian Ocean and western North Pacific (Fig. 10a) results from the increased AA  
415 advected by prevailing winds from South and East Asia. This cooling corresponds to  
416 the decrease of water vapor extending from western North Pacific to East Asia (Fig.  
417 10b). The decrease of water vapor in the atmosphere results in decrease of cloud cover  
418 over Eastern China (Fig. 10c), which induces positive SW CRE (Fig. 10d). The  
419 positive SW CRE, with a magnitude of  $4.01 \text{ W m}^{-2}$  over NC and SEC, contributes to  
420 the local surface warming (Fig. 10a) and decrease in cold extremes over NC and SEC.  
421 In addition, the decrease of upward latent heat flux (Fig. 10e;  $1.33 \text{ W m}^{-2}$ ), as a  
422 consequences of decrease in evaporation (Fig. 10f), also makes a contribution to the  
423 surface warming and a contribution to increase in TXn and TNn and decrease in  
424 frequencies of ID and FD over SEC.

## 425 **7. Conclusions**

426 We found significant decadal changes in both hot and cold extremes over China  
427 since the mid-1990s by using Chinese observed station dataset. These changes are  
428 characterized as the rise in TXx, TNx and the increase in frequencies of SU and TR  
429 during summer, and the rise in TXn and TNn and the decrease in frequencies of ID and  
430 FD during winter. In this study, we have performed a set of experiments using an  
431 atmosphere-ocean-mixed-layer coupled model to assess the contributions of All forcing  
432 changes, as an combined effects of GHG and AA forcing, to observed decadal changes  
433 in temperature extremes over China across the mid-1990s, and quantify the relatively  
434 roles of changes in GHG concentrations and AA emissions, respectively. The main



435 conclusions are as follow.

436 Observations indicate that there was an abrupt change in temperature extremes over  
437 China since the mid-1990s. The changes of temperature extremes are analyzed by the  
438 comparison between the PD of 1994~2011 and the EP of 1964~1981. Spatially  
439 averaged over China, the hot extremes of TXx and TNx are increased by 0.58 °C and  
440 0.76 °C, respectively. The frequencies of SU and TR are increased by about 7~9 days.  
441 The cold extremes of TXn and TNn are increased by 1.48 °C and 1.82 °C, respectively.  
442 The frequencies of ID and FD are decreased by about one week. Furthermore, these  
443 abrupt decadal changes occur not only over China as a whole, but also over three sub-  
444 regions of NC, SEC and SWC, even though they exhibit various climate types.

445 The atmosphere-ocean-mixed-layer coupled model MetUM-GOML1 in response  
446 to changes in GHG concentrations and AA emissions together (All forcing) realistically  
447 reproduces the spatial patterns of the observed decadal changes in both hot and cold  
448 temperature extremes. Quantitatively, the model simulated changes in response to All  
449 forcing changes are comparable to the observed decadal changes. The results indicate  
450 a dominant role of anthropogenic changes in the observed decadal changes of  
451 temperature extremes over China across the mid-1990s.

452 Moreover, model responses to changes in GHG concentration and AA emissions  
453 show some different characteristics. GHG forcing changes lead to increase in hot  
454 extremes (TXx, TNx, SU and TR), and TNx and TNn and decrease in frequencies of  
455 ID and FD over China, while AA forcing changes lead to weak increases in hot extremes

456 over NC and decrease over SEC during summer and induce changes in cold extremes  
457 the same sign as those induced by GHG over NC and SEC during winter, but with weak  
458 magnitude. The responses of cold extremes in response to changes in GHG forcing are  
459 two to three times as large as those in response to changes in AA forcing, indicating a  
460 dominant role of GHG forcing changes in the model simulated cold extreme changes  
461 in response to All forcing changes. Relatively, the increase of extremes over NC in  
462 responses to changes in GHG forcing explains up to 75% of the TNx, 90% of the TXn  
463 and 60% of the TNn increase in response to changes in All forcing (assuming linearity).  
464 These results indicate that the model simulated extreme temperature changes in  
465 response to All forcing changes are predominantly induced by GHG forcing change,  
466 but AA change also makes some weak contributions.

467 In response to the increase of GHG concentrations, the increase of hot extremes is  
468 mainly due to the increased Greenhouse Effect with positive net surface clear sky LW  
469 radiation. Additionally, the increase of net surface SW radiation, mainly resulted from  
470 the positive SW CRE, associated with the decrease of cloud cover, has a positive  
471 feedback with the surface warming and the increase in hot extremes. In terms of  
472 changes in cold extremes in response to GHG forcing changes, the increase of net  
473 surface clear sky LW radiation due to the increased greenhouse effect and associated  
474 water vapor feedback tend to result in surface warming over southern part of China,  
475 and therefore lead to increase in TXn and TNn and decrease in frequencies of ID and  
476 FD. The changes of cold day-time extremes (TXn and ID) over NC are further enhanced

477 by the increase of clear sky SW radiation related to the decrease of snow-albedo  
478 feedback.

479 During summer, the response of hot extremes over China to changes in AA forcing  
480 exhibits a dipole pattern with increases in north and decreases in south. Local increase  
481 of AOD over East Asia leads to decrease of net clear sky SW radiation, which tends to  
482 cool the surface. However, the positive SW CRE tends to warm the surface and leads  
483 to increase in hot extremes over NC. The positive SW CRE is induced by the decrease  
484 of cloud cover, which related to the decrease of soil moisture, as a consequence of  
485 reduction in precipitation over NC, while decreases in hot extremes over South China  
486 are the results of direct impacts of AA forcing changes through aerosol-radiation and  
487 aerosol-cloud interactions due to increased AA emissions over East Asia. During winter,  
488 AA are advected by mean flow to the Indian Ocean and western North Pacific, which  
489 induces cooling over there. This cooling reduces water vapor and therefore reduces  
490 cloud cover over East Asia, leading to positive SW CRE over NC and SEC and  
491 therefore leading to increase in TXn and TNn and decrease in frequencies of ID and FD  
492 over NC and SEC.

493 In this study, besides the cooling effect by the direct impacts of AA forcing changes  
494 through aerosol-radiation and aerosol-cloud interactions, we find a surface warming  
495 over NC during summer and over China during winter driven by AA forcing changes  
496 through the AA induced atmosphere-cloud feedback. This aerosol-climate interaction  
497 is consistent with Tian et al. (2018)..

498 Our results suggested the different roles of GHG and AA in temperature extremes.  
499 The model shows a strong warming over China in response to GHG forcing, and a  
500 cooling over SC and a weak warming over NC in response to AA forcing during  
501 summer. In addition, previous studies have pointed out different roles of GHG and AA  
502 in shaping the temperature trend over China by using CMIP5 models. The CMIP5  
503 experiments result suggested that the GHG plays a dominant role in the warming trend  
504 over China (e.g., Song et al. 2014; Zhao et al. 2015), which is consistent with our model  
505 result. Zhao et al. (2015) showed that the AA forcing has a cooling effect, and they  
506 further indicated that the individual effects of AA cannot be detected in the observed  
507 temperature changes with respect to the combined effects among all the other forcings,  
508 implying an uncertainty about the AA forcing impact in their study. In addition, Li et  
509 al. (2015) further suggested that the indirect AA effect (including indirect, semidirect,  
510 surface albedo effects, and so on) induce warming over NC, which also can be seen by our  
511 model result.

512 The results in this study indicate a remarkable role of anthropogenic changes,  
513 especially the increased GHG concentrations, in the observed decadal changes of  
514 temperature extremes over China since the mid-1990s. Given the fact that GHG  
515 concentrations and local AA emissions will continue to rise in the next few decades,  
516 observed recent decadal changes in temperature extremes over China are likely to  
517 sustain, or even amplify in the near future.

518

519 **Acknowledgement**

520 This study is supported by the National Natural Science Foundation of China under  
521 Grants 41675078, U1502233, 41320104007, by the Youth Innovation Promotion  
522 Association of CAS (No. 2018102) and by the UK-China Research & Innovation  
523 Partnership Fund through the Met Office Climate Science for Service Partnership  
524 (CSSP) China as part of the Newton Fund. BD is supported by the U.K. National Centre  
525 for Atmospheric Science–Climate (NCAS-Climate) at the University of Reading. The  
526 authors thank Editor Jian Lu and anonymous reviewers for their constructive comments  
527 on the earlier version of the paper.

528 **Reference**

- 529 Alexander LV et al (2006) Global observed changes in daily climate extremes of  
530 temperature and precipitation. *J Geophys Res* 111:D05109. doi:  
531 10.1029/2005JD006290
- 532 Arribas A et al (2011) The GloSea4 ensemble prediction system for seasonal forecasting.  
533 *Monthly Weather Review* 139(6):1891–1910
- 534 Barsugli J, Battisti D S (1998) The basic effects of atmosphere–ocean thermal coupling  
535 on midlatitude variability. *J Atmos Sci* 55:477–493. doi:10.1175/1520-0469
- 536 Bellouin N, Rae J, Jones A, Johnson C, Haywood J, Boucher O (2011) Aerosol forcing  
537 in the climate model intercomparison project (CMIP5) simulations by  
538 HadGEM2-ES and the role of ammonium nitrate. *J Geophys Res* 116:D20206.  
539 doi: 10.1029/2011JD016074
- 540 Boé J, Terray L (2014) Land–sea contrast, soil-atmosphere and cloud-temperature  
541 interactions: interplays and roles in future summer European climate change.  
542 *Clim Dyn* 42:683–699
- 543 Bollasina MA, Ming Y, Ramaswamy V (2011) Anthropogenic aerosols and the  
544 weakening of the South Asian summer monsoon. *Science* 334(6055):502–505
- 545 Bony S et al (2006) How well do we understand and evaluate climate change feedback  
546 processes? *J Clim* 19(15):3445–3482. doi: 10.1175/jcli3819.1
- 547 Christidis N et al (2013) A new HadGEM3-A-based system for attribution of weather  
548 and climate-related extreme events. *J Clim* 26:2756–2783. doi:10.1175/JCLI-D-

549 12-00169.1

550 Cubasch U et al (2001) Projections of future climate change. *Climate Change 2001:*  
551 *The Scientific Basis*, J. T. Houghton et al., Eds., Cambridge University Press  
552 525–582

553 Donat MG, Alexander LV, Yang H, Durre I, Vose R, Caesar J (2013) Global land-based  
554 datasets for monitoring climatic extremes. *Bull Am Meteorol Soc* 94:997–1006.  
555 doi:10.1175/BAMS-D-12-00109.1

556 Dong BW, Sutton RT, Shaffrey L (2017a) Understanding the rapid summer warming  
557 and changes in temperature extremes since the mid-1990s over Western Europe.  
558 *Clim Dyn* 48:1537–1554. doi:10.1007/s00382-016-3158-8

559 Dong BW, Sutton RT, Shaffrey L, Klingaman NP (2017b) Attribution of forced decadal  
560 climate change in coupled and uncoupled ocean-atmosphere model experiments.  
561 *J Clim* doi:10.1175/JCLI-D-16-0578.1

562 Dong BW, Sutton RT, Chen W, Liu XD, Lu RY, Sun Y (2016) Abrupt summer warming  
563 and changes in temperature extremes over Northeast Asia since the mid-1990s:  
564 Drivers and physical processes. *Adv Atmos Sci* 33(9):1005–1023. doi:  
565 10.1007/s00376-016-5247-3

566 Dong BW, Sutton RT, Highwood E, Wilcox L (2015) Preferred response of the East  
567 Asian summer monsoon to local and nonlocal anthropogenic sulphur dioxide  
568 emissions. *Clim Dyn* doi:10.1007/s00382-015-2671-5

569 Dong BW, Gregory JM, Sutton RT (2009) Understanding land-sea warming contrast in

570 response to increasing greenhouse gases. Part I: Transient adjustment. *J Clim*  
571 22:3079-3097

572 Dwyer JG, Biasutti M, Sobel AH (2012) Projected changes in the seasonal cycle  
573 of surface temperature. *J Clim* 25(18):6359–6374

574 Feichter J, Roeckner E, Lohmann U, Liepert B (2004) Nonlinear aspects of the  
575 climate response to greenhouse gas and aerosol forcing. *J Clim* 17:2384–  
576 2398

577 Freychet S, Tett S, Wang J, Hegerl G (2017) Summer heat waves over eastern china:  
578 dynamical processes and trend attribution. *Environ Res Lett* 12:1–9. doi:  
579 10.1088/1748-9326/aa5ba3

580 Hansen J, Sato M, Ruedy R (1997) Radiative forcing and climate response. *J Geophys*  
581 *Res* 102:6831–6864. doi:10.1029/96JD03436

582 He J, Soden B (2016) Does the lack of coupling in SST-forced atmosphere-only models  
583 limit their usefulness for climate change studies? *J Clim* 29:4317–4325.  
584 doi:10.1175/JCLI-D-14-00597.1

585 Hiron LC, Klingaman NP, Woolnough SJ (2015) MetUM-GOML: a near-globally  
586 coupled atmosphere–ocean-mixed-layer model. *Geoscientific Model*  
587 *Development* 8:363–379

588 Jones C et al (2011) The HadGEM2-ES implementation of CMIP5 centennial  
589 simulations. *Geophys Model Dev* 4:543–570

590 Kamae Y, Shiogama H, Watanabe M, Kimoto M (2014) Attributing the increase in



591 Northern Hemisphere hot summers since the late 20th century. *Geophys Res Lett*  
592 41:5192–5199. doi:10.1002/2014GL061062

593 Kim YH, Min SK, Zhang X, Zwiers F, Alexander LV, Donat MG, Tung YS (2015)  
594 Attribution of extreme temperature changes during 1951–2010. *Clim Dyn*  
595 46:1769–1782. doi:10.1007/s00382-015-2674-2

596 Lamarque JF et al (2010) Historical (1850–2000) gridded anthropogenic and biomass  
597 burning emissions of reactive gases and aerosols: Methodology and application.  
598 *Atmos Chem Phys* 10:7017–7039. doi:10.5194/acp-10-7017-2010

599 Lamarque JF et al (2011) Global and regional evolution of short-lived radiatively-active  
600 gases and aerosols in the Representative Concentration Pathways. *Climatic*  
601 *Change* 109:191–212. doi:10.1007/s10584-011-0155-0

602 Li CX, Zhao TB, Ying KR (2015) Effects of anthropogenic aerosols on temperature  
603 changes in China during the twentieth century based on CMIP5 models. *Theor*  
604 *Appl Climatol* doi: 10.1007/s00704-015-1527-6

605 Li X, Ting M (2016) Understanding the Asian summer monsoon response to  
606 greenhouse warming: the relative roles of direct radiative forcing and sea surface  
607 temperature change. *Clim Dyn* 49:1–18

608 Li Z et al (2016) Comparison of two homogenized datasets of daily  
609 maximum/mean/minimum temperature in China during 1960–2013. *J Meteor*  
610 *Res* 30(1):053–066. doi: 10.1007/s13351-016-5054-x

611 Ma SM, Zhou TJ, Stone D, Angelil O, Shiogama H (2017) Attribution of the July-

612 August 2013 heat event in central and eastern China to anthropogenic  
613 Greenhouse gas emissions. *Environ Res Lett* 12:054020

614 Ming Y, Ramaswamy V (2009) Nonlinear climate and hydrological responses to  
615 aerosol effects. *J Clim* 22:1329–1339

616 Otto FE, Massey, van Oldenborgh GJ, Jones RQ, Allen MR (2012) Reconciling two  
617 approaches to attribution of the 2010 Russian heat wave. *Geophys Res Lett*  
618 39:L04702. doi:10.1029/2011GL050422

619 Perkins SE (2015) A review on the scientific understanding of heatwaves--their  
620 measurement, driving mechanisms, and changes at the global scale. *Atmos Res*  
621 164:242–267

622 Qu X, Hall A (2007) What controls the strength of snow-albedo feedback? *J Clim*  
623 20(15):3971–3981, doi: doi:10.1175/JCLI4186.1

624 Randall DA et al (1994) Analysis of snow feedbacks in 14 general circulation models.  
625 *J Geophys Res Atmos* 99(D10):20757–20771. doi: 10.1029/94jd01633

626 Rangwala I, Sinsky E, Miller JR (2016) Variability in projected elevation dependent  
627 warming in boreal midlatitude winter in CMIP5 climate models and its potential  
628 drivers. *Clim Dyn* 46:2115

629 Rayner NA et al (2003) Global analyses of sea surface temperature, sea ice, and night  
630 marine air temperature since the late nineteenth century. *J Geophys Res* 108:4407.  
631 doi:10.1029/2002JD002670

632 Robock A (1983) Ice and snow feedbacks and the latitudinal and seasonal distribution

633 of climate sensitivity. *J Atmos Sci* 40 (4):986–997

634 Schaller N et al (2016) Human influence on climate in the 2014 southern England  
635 winter floods and their impacts. *Nat Clim Change* 6:627–634.  
636 doi:10.1038/nclimate2927

637 Shiogama H, Stone DA, Nagashima T, Nozawa T, Emori S (2012) On the linear  
638 additivity of climate forcing-response relationships at global and  
639 continental scales. *Int J Climatol* 33:2542–50

640 Smith DM, Murphy JM (2007) An objective ocean temperature and salinity analysis  
641 using covariances from a global climate model. *J Geophys Res* 112:C02022.  
642 doi:10.1029/2005JC003172

643 Song F, Zhou T, Qian Y (2014) Responses of East Asian summer monsoon to natural  
644 and anthropogenic forcings in the 17 latest CMIP5 models. *Geophys Res Lett* 41.  
645 doi:10.1002/2013GL058705

646 Stevens B, Feingold G (2009) Untangling aerosol effects on clouds and precipitation in  
647 a buffered system. *Nat* 461:607–613. doi:10.1038/nature08281

648 Stouffer RJ, Wetherald RT (2007) Changes of variability in response to increasing  
649 greenhouse gases. part i: temperature. *J Clim* 20(21):5455

650 Stott et al (2016) Attribution of extreme weather and climate-related events. Wiley  
651 *Interdiscip. Rev Clim Change* 7:23–41. doi:10.1002/wcc.380

652 Sun Y, Song LC, Yin H, Zhang XB, Stott P, Zhou BT, Hu T (2016) Human Influence  
653 on the 2015 extreme high temperature events in western China [in “Explaining

654 Extreme Events of 2015 from a Climate Perspective”]. Bull Amer Meteor Soc  
655 97:S5–S9.

656 Sun Y, Zhang X, Zwiers FW, Song L, Wan H, Hu T, Yin H, Ren G (2014) Rapid increase  
657 in the risk of extreme summer heat in Eastern China. Nat Clim Change 4:1082–  
658 1085. doi:10.1038/nclimate2410

659 Tian FX, Dong BW, Robson J Sutton RT (2018) Forced decadal changes in the East  
660 Asian summer monsoon: the roles of greenhouse gases and anthropogenic  
661 aerosols. Clim Dyn 6:1–17

662 Thackeray CW, Fletcher CG (2016) Snow albedo feedback: Current knowledge,  
663 importance, outstanding issues and future directions. Prog Phys Geogr  
664 40(3):392–408. doi: 10.1177/0309133315620999

665 Vanni re BE, Guilyardi G, Madec FJ, Dobl s R, Woolnough S (2013) Using seasonal  
666 hindcasts to understand the origin of the equatorial cold tongue bias in CGCMs  
667 and its impact on ENSO. Clim Dyn 40(3–4):963–981

668 Walters DN, Best MJ, Bushell AC, Copsey D, Edwards JM, Falloon PD, Roberts MJ  
669 (2011) The met office unified model global atmosphere 3.0/3.1 and JULES  
670 global land 3.0/3.1 configurations. Geosci Model Dev 4(4):919

671 Wang T et al (2013) Anthropogenic agent implicated as a prime driver of shift in  
672 precipitation in eastern China in the late 1970s. Atmos Chem Phys 13(24):12433

673 Wang T, Otter  OH, Gao YG, Wang HJ (2012) The response of the North Pacific  
674 Decadal Variability to strong tropical volcanic eruptions. Clim Dyn 39(12):2917–

675           2936

676   Wei K, Chen W (2011) An abrupt increase in the summer high temperature extreme  
677           days across China in the mid-1990s. *Adv Atmos Sci* 28(5):1023–1029. doi:  
678           10.1007/s00376-010-0080-6

679   Wen HQ, Zhang X, Xu Y, Wang B (2013) Detecting human influence on extreme  
680           temperatures in China. *Geophys Res Lett* 40:1171–1176. doi:10.1002/grl.50285.

681   Wilcox LJ, Dong BW, Sutton RT, Highwood EJ (2015) The 2014 Hot, Dry Summer in  
682           Northeast Asia [in “Explaining Extreme Events of 2014 from a Climate  
683           Perspective”]. *Bull Amer Meteor Soc* 96(12):S105–S110. doi:10.1175/BAMS-  
684           D-15-00123.1

685   Yang FL et al (2001) Snow-albedo feedback and seasonal climate variability over North  
686           America. *J Clim* 14(22):4245–4248

687   Yang Y, Russell LM, Lou S, Lamjiri MA, Liu Y, Singh B, Ghan SJ (2016) Changes in  
688           sea salt emissions enhance ENSO variability, *J Clim* 29:8575–8588.  
689           doi:10.1175/JCLI-D-16-0237.1

690   Yang Y, Russell LM, Lou S, Liao H, Guo J, Liu Y, Singh B, Ghan SJ (2017) Dust-wind  
691           interactions can intensify aerosol pollution over eastern China. *Nat Commun*  
692           8:15333. doi:10.1038/ncomms15333

693   Yao Y, Luo Y, Huang JB (2012) Evaluation and projection of temperature extremes over  
694           China based on CMIP5 model. *Adv Clim Change Res* 3(4). doi:  
695           10.3724/SP.J.1248.2012.00179

696 Yin H, Sun Y, Wan H, Zhang XB, Lu CH (2016) Detection of anthropogenic influence  
697 on the intensity of extreme temperatures in China. *Int J Climatol* 37:1229–1237.  
698 doi: 10.1002/joc.4771

699 Zhao TB, Li CX, Zuo ZY (2016) Contributions of anthropogenic and external natural  
700 forcings to climate changes over China based on CMIP5 model simulations. *Sci*  
701 *China Earth Sci* 59:503–517. doi: 10.1007/s11430-015-5207-2

702 Zhou BT, Xu Y, Wu J, Dong S, Shi Y (2016) Changes in temperature and precipitation  
703 extreme indices over China: analysis of a high-resolution grid dataset. *Int J*  
704 *Climatol* 36:1051–1066

705 **Figure captions**

706 **Table 1.** Summary of numerical experiments.

707 **Figure 1.** Time series of annual mean temperature extremes anomalies relative to the  
708 climatology (mean of the whole period) in summer (TX<sub>x</sub>, TN<sub>x</sub>, SU and TR; left  
709 panels) and in winter (TX<sub>n</sub>, TN<sub>n</sub>, ID and FD; right panels) over China by GHCND  
710 dataset (averaged over 20°~55°N, 75°~130°E; black solid lines) and by China  
711 station dataset (averaged over 753 stations; red solid lines). The color dashed lines  
712 represent the time series of temperature extremes anomalies averaged over three  
713 sub-regions by China station dataset (see their distributions in Fig. 2a). Black  
714 dashed range bars indicate the early period (EP) of 1964~1981 and the present day  
715 (PD) of 1994~2011. Units in TX<sub>x</sub>, TN<sub>x</sub>, TX<sub>n</sub>, and TN<sub>n</sub> are °C. Units in SU, TR,  
716 ID and FD are days.

717 **Figure 2.** (a) Distributions of 753 stations in China station dataset. The three sub-  
718 regional groups are marked with different color dots. The dots in blue, purple and  
719 green represent the sub-regions of Northern China (NC), Southeastern China (SEC)  
720 and Southwestern China (SWC), respectively. (b)-(i) Spatial patterns of differences  
721 in temperature extremes in summer (left panels) and in winter (right panels)  
722 between the PD and the EP. The black lines indicate the regions where the changes  
723 are statistically significant at the 90% confidence level base on *t*-test. Units in TX<sub>x</sub>,  
724 TN<sub>x</sub>, TX<sub>n</sub> and TN<sub>n</sub> are °C. Units in SU, TR, ID and FD are days.

725 **Figure 3.** Differences in annual mean sulfur dioxide emissions (units: g m<sup>-2</sup> yr<sup>-1</sup>)  
726 between 1994~2010 and 1970~1981.

727 **Figure 4.** Spatial patterns of changes in hot extremes in response to changes in All  
728 forcing (left panels), GHG forcing (middle panels) and AA forcing (right panels),  
729 being masked by China boundary. The black lines indicate the regions where the  
730 changes are statistically significant at the 90% confidence level base on *t*-test. Units  
731 in TX<sub>x</sub> and TN<sub>x</sub> are °C. Units in SU and TR are days.

732 **Figure 5.** Same as Fig. 4, but for changes in cold extremes. Units in TN<sub>x</sub> and TN<sub>n</sub>  
733 are °C. Units in ID and FD are days.

734 **Figure 6.** Observed and model simulated changes in temperature extremes in response  
735 to different forcings over Northern China (a, b; NC, 35°~55°N, 75°~130°E),  
736 Southeastern China (c, d; SEC; 20°~35°N, 105°~130°E) and Southwestern China  
737 (e, f; SWC; 20°~35°N, 75°~105°E). The model simulated values have been masked  
738 by China boundary. The color bars indicate central estimates and dots show the 90%  
739 confidence intervals based on two-tailed Students' *t*-test. Top panels for TX<sub>x</sub>, TX<sub>n</sub>,  
740 TN<sub>x</sub> and TN<sub>n</sub> (units: °C) and bottom panels for SU, TR, ID and FD (units: days).

741 **Figure 7.** Spatial patterns of summer mean response to changes in GHG forcing: (a)  
742 clear sky LW radiation; (b) surface LW radiation; (c) total cloud cover (units: %)  
743 (d) low level cloud cover (units: %); (e) SW CRE; and (f) surface SW radiation.  
744 Radiation is the net component and in W m<sup>-2</sup> with positive value meaning  
745 downward. The black lines highlight regions where the changes are statistically  
746 significant at the 90% confidence level base on *t*-test.

747 **Figure 8.** Spatial patterns of summer mean response to changes in AA forcing: (a) total  
748 AOD; (b) clear sky SW radiation; (c) SW CRE; (d) total cloud cover (units: %); (e)  
749 soil moisture (units: kg m<sup>-2</sup>); and (f) column-integrated water vapor (units: kg m<sup>-2</sup>).  
750 Radiation is the net component and in W m<sup>-2</sup> with positive value meaning  
751 downward. The black lines highlight regions where the changes are statistically  
752 significant at the 90% confidence level base on *t*-test.

753 **Figure 9.** Spatial patterns of winter mean response to changes in GHG forcing: (a) clear  
754 sky LW radiation; (b) surface LW radiation; (c) column-integrated water vapor  
755 (units: kg m<sup>-2</sup>); (d) surface SW radiation; (e) clear sky SW radiation; and (f) skin  
756 temperature (units: °C). Radiation is the net component and in W m<sup>-2</sup> with positive  
757 value meaning downward. The black lines highlight regions where the changes are  
758 statistically significant at the 90% confidence level base on *t*-test.

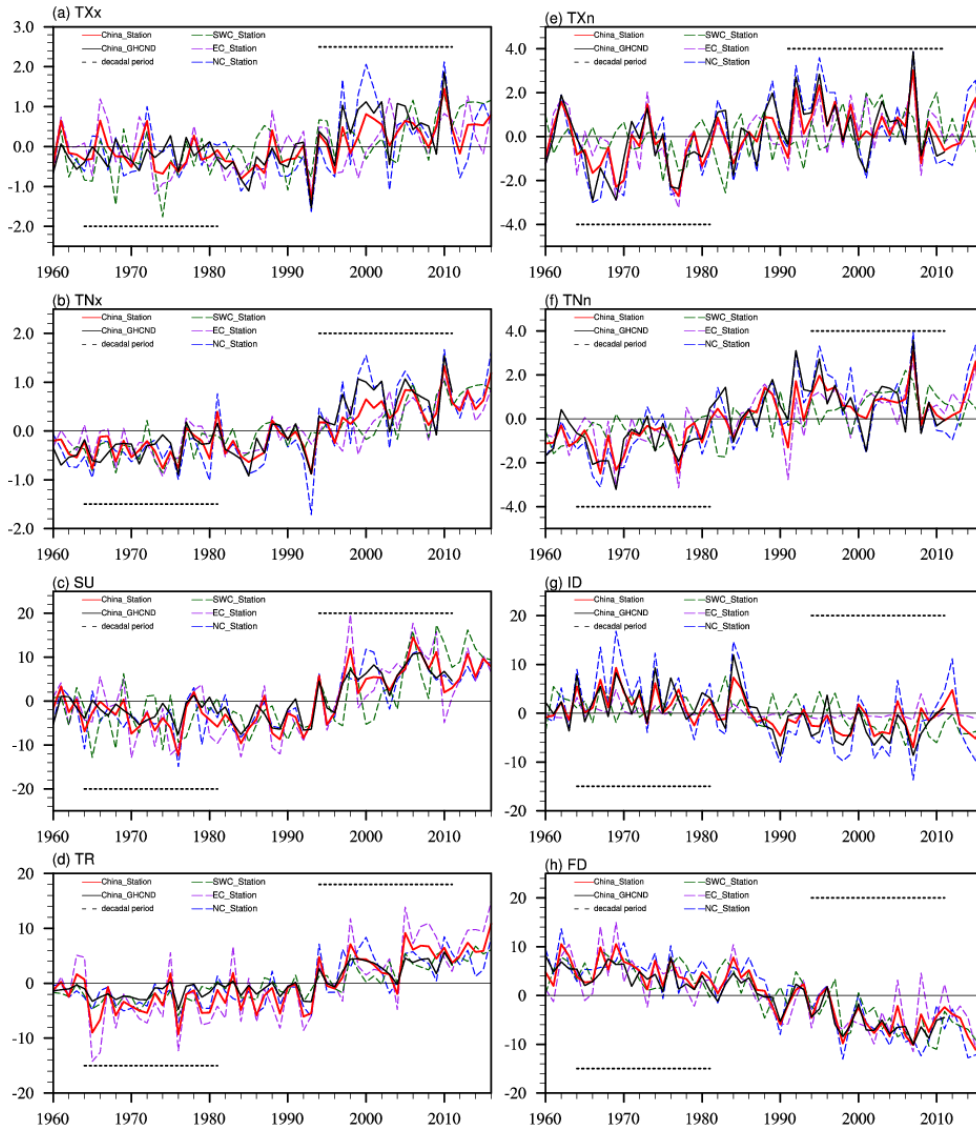
759 **Figure 10.** Spatial patterns of winter mean response to changes in AA forcing: (a) skin



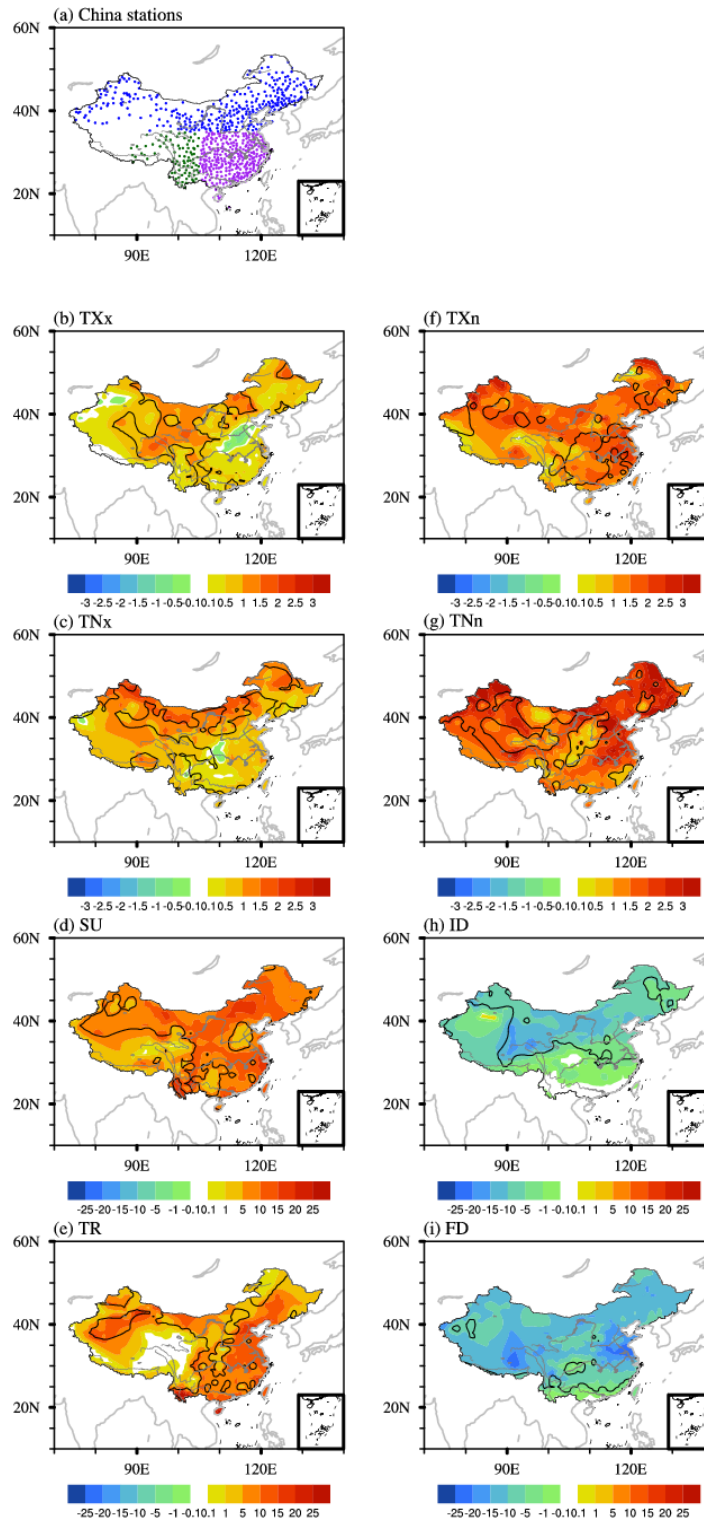
760 temperature (units: °C); (b) column-integrated water vapor (units: kg m<sup>-2</sup>); (c) total  
761 cloud cover (units: %); (d) SW CRE; (e) surface latent heat flux; and (f)  
762 evaporation (units: kg m<sup>-2</sup>). Radiation and flux are in W m<sup>-2</sup> with positive value  
763 meaning downward. The black lines highlight regions where the changes are  
764 statistically significant at the 90% confidence level based on *t*-test.

**Table 1.** Summary of numerical experiments. Note that a slightly different period of 1970–1981 for the aerosol forcing in the early period is used since aerosol emissions data before 1970 were not available

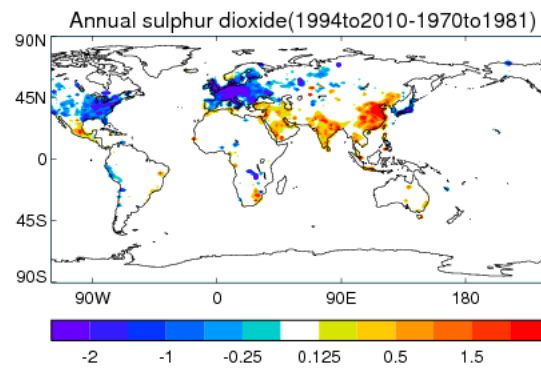
Adv.	Experiment	Ocean	Radiative Forcing
<b>R0</b>	Relaxation run	Relaxation to “present day” (PD, 1994-2011) mean 3D ocean temperature and salinity to diagnose climatological temperature and salinity tendencies	PD greenhouse gases (GHGs) over 1994~2011 and anthropogenic aerosol (AA) emissions over 1994~2010 with AA after 2006 from RCP4.5 scenario (Lamarque et al. 2010, 2011)
<b>EP</b>	Early period (EP 1964~1981)	Climatological temperature and salinity tendencies from relaxation run	EP GHGs over 1964~1981 and AA emissions over 1970~1981
<b>PDGA</b>	Present Day (PD 1994~2011) with GHG and AA forcings	Climatological	PD GHG and PD AA emissions
<b>PDG</b>	Present Day (PD 1994~2011) with GHG forcing	temperature and salinity tendencies from relaxation run	PD GHG and EP AA emissions
<b>PDA</b>	Present Day (PD 1994~2011) with AA forcing		EP GHG and PD AA emissions



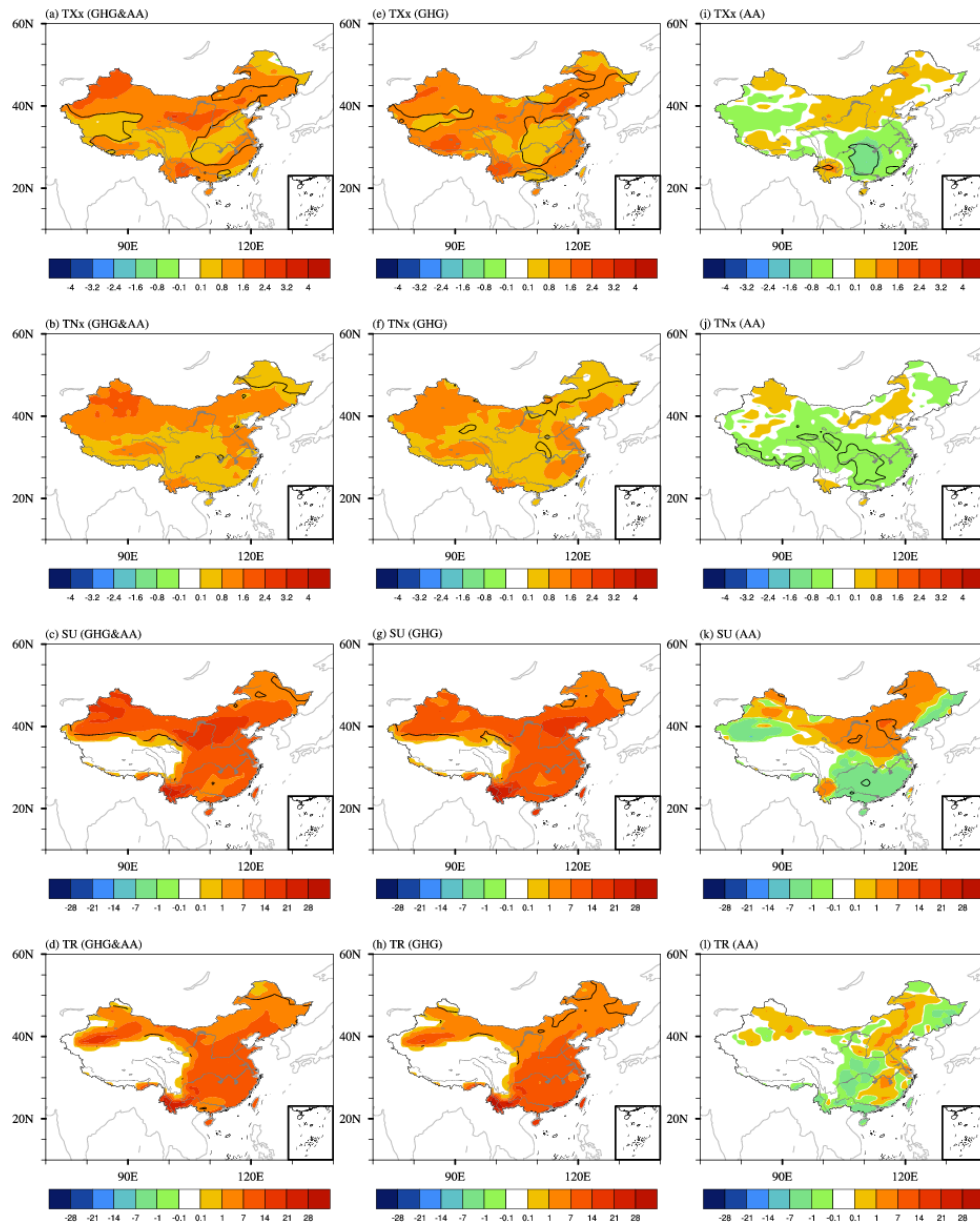
**Figure 1.** Time series of annual mean temperature extremes anomalies relative to the climatology (mean of the whole period) in summer (TX<sub>x</sub>, TN<sub>x</sub>, SU and TR; left panels) and in winter (TX<sub>n</sub>, TN<sub>n</sub>, ID and FD; right panels) over China by GHCND dataset (averaged over 20°~55°N, 75°~130°E; black solid lines) and by China station dataset (averaged over 753 stations; red solid lines). The color dashed lines represent the time series of temperature extremes anomalies averaged over three sub-regions by China station dataset (see their distributions in Fig. 2a). Black dashed range bars indicate the early period (EP) of 1964~1981 and the present day (PD) of 1994~2011. Units in TX<sub>x</sub>, TN<sub>x</sub>, TX<sub>n</sub>, and TN<sub>n</sub> are °C. Units in SU, TR, ID and FD are days.



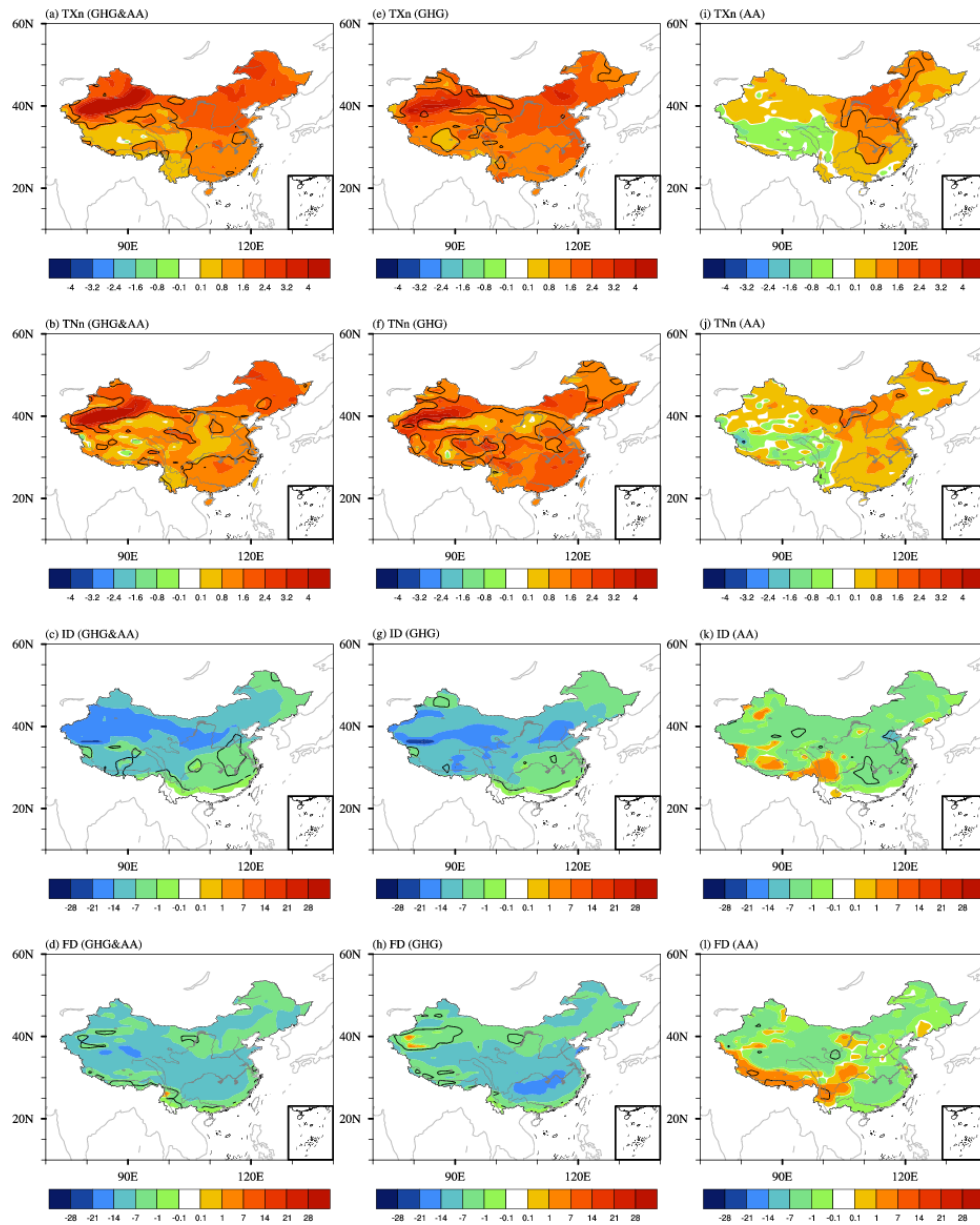
**Figure 2.** (a) Distributions of 753 stations in China station dataset. The three sub-regional groups are marked with different color dots. The dots in blue, purple and green represent the sub-regions of Northern China (NC), Southeastern China (SEC) and Southwestern China (SWC), respectively. (b)-(i) Spatial patterns of differences in temperature extremes in summer (left panels) and in winter (right panels) between the PD and the EP. The black lines indicate the regions where the changes are statistically significant at the 90% confidence level base on  $t$ -test. Units in TXx, TNx, TXn and TNn are  $^{\circ}\text{C}$ . Units in SU, TR, ID and FD are days.



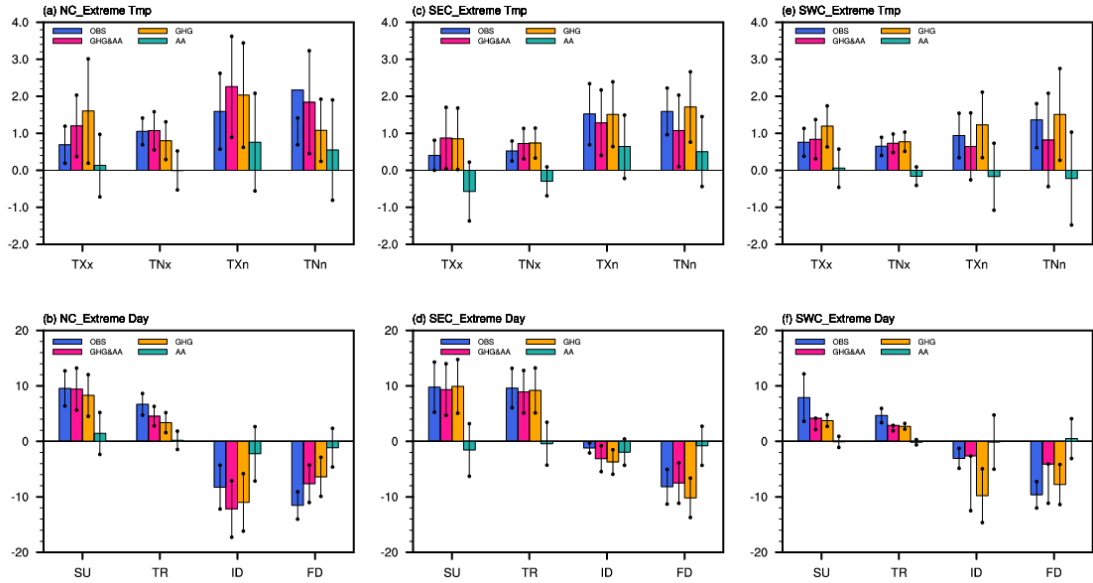
**Figure 3.** Differences in annual mean sulfur dioxide emissions (units: g m<sup>-2</sup> yr<sup>-1</sup>) between 1994~2010 and 1970~1981.



**Figure 4.** Spatial patterns of changes in hot extremes in response to changes in All forcing (left panels), GHG forcing (middle panels) and AA forcing (right panels), being masked by China boundary. The black lines indicate the regions where the changes are statistically significant at the 90% confidence level base on  $t$ -test. Units in TXx and TNx are °C. Units in SU and TR are days.

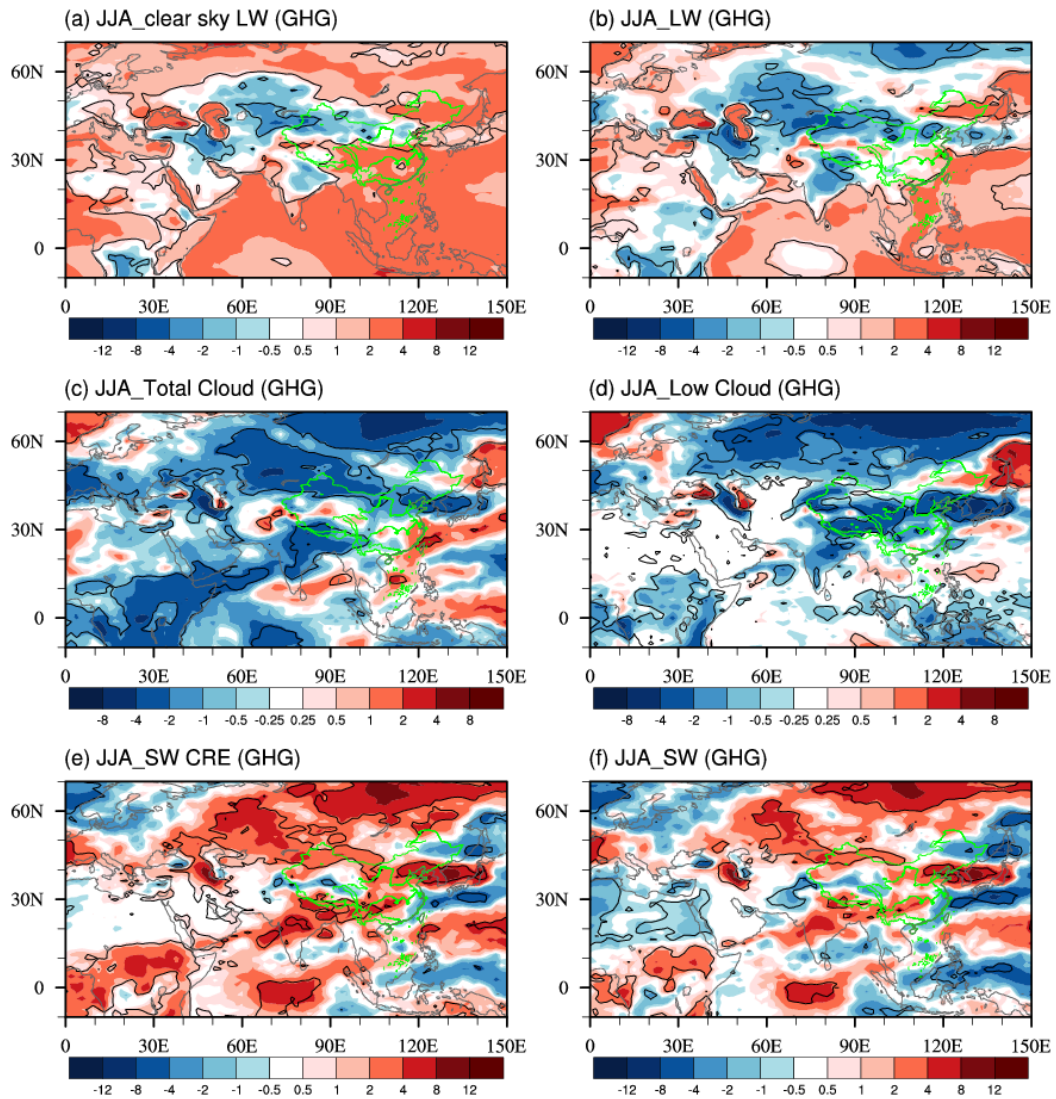


**Figure 5.** Same as Fig. 4, but for changes in cold extremes. Units in TNx and TNn are  $^{\circ}\text{C}$ . Units in ID and FD are days.

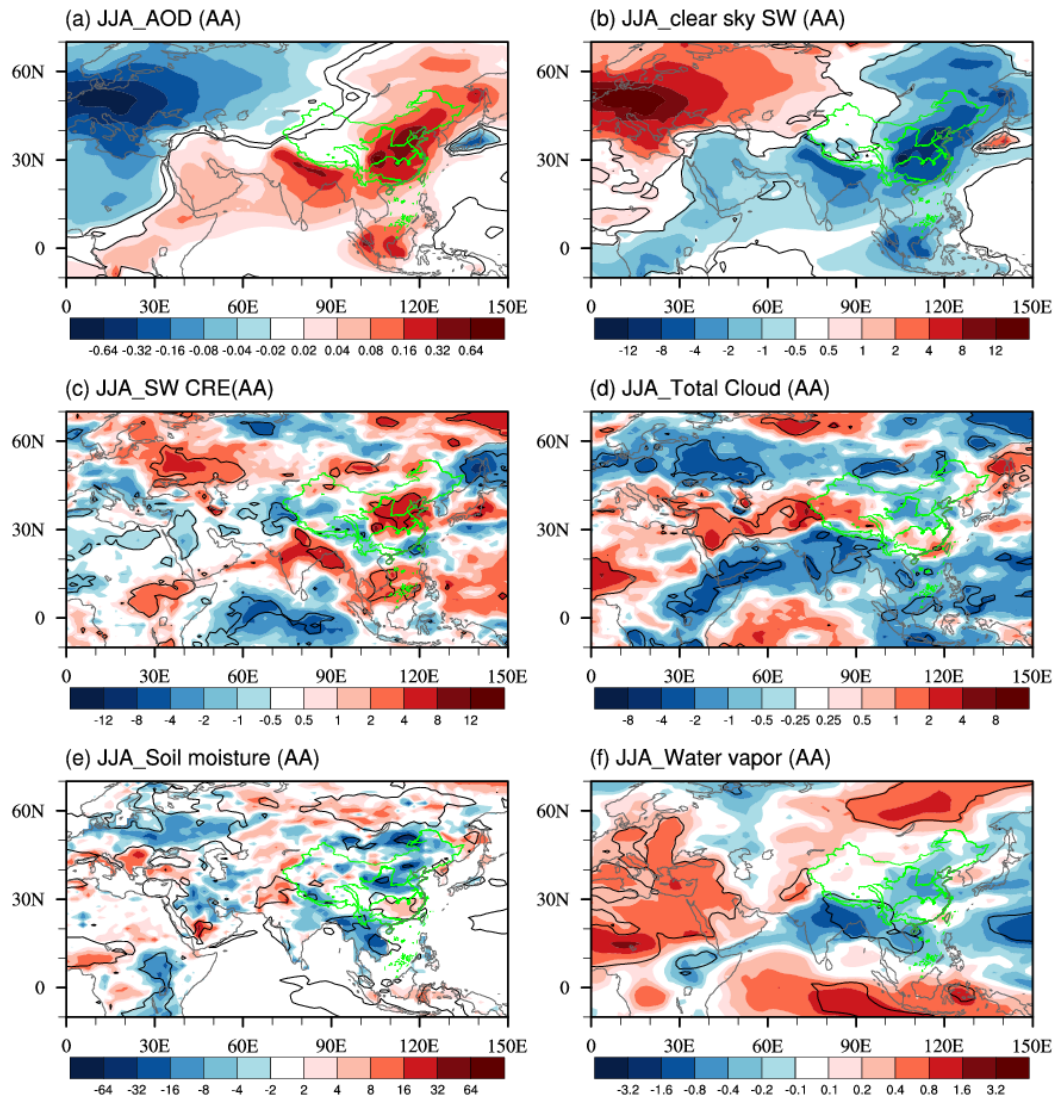


**Figure 6.** Observed and model simulated changes in temperature extremes in response to different forcings over Northern China (a, b; NC, 35°~55°N, 75°~130°E), Southeastern China (c, d; SEC; 20°~35°N, 105°~130°E) and Southwestern China (e, f; SWC; 20°~35°N, 75°~105°E). The model simulated values have been masked by China boundary. The color bars indicate central estimates and dots show the 90% confidence intervals based on two-tailed Students' *t*-test. Top panels for TXx, TXn, TNx and TNn (units: °C) and bottom panels for SU, TR, ID and FD (units: days).

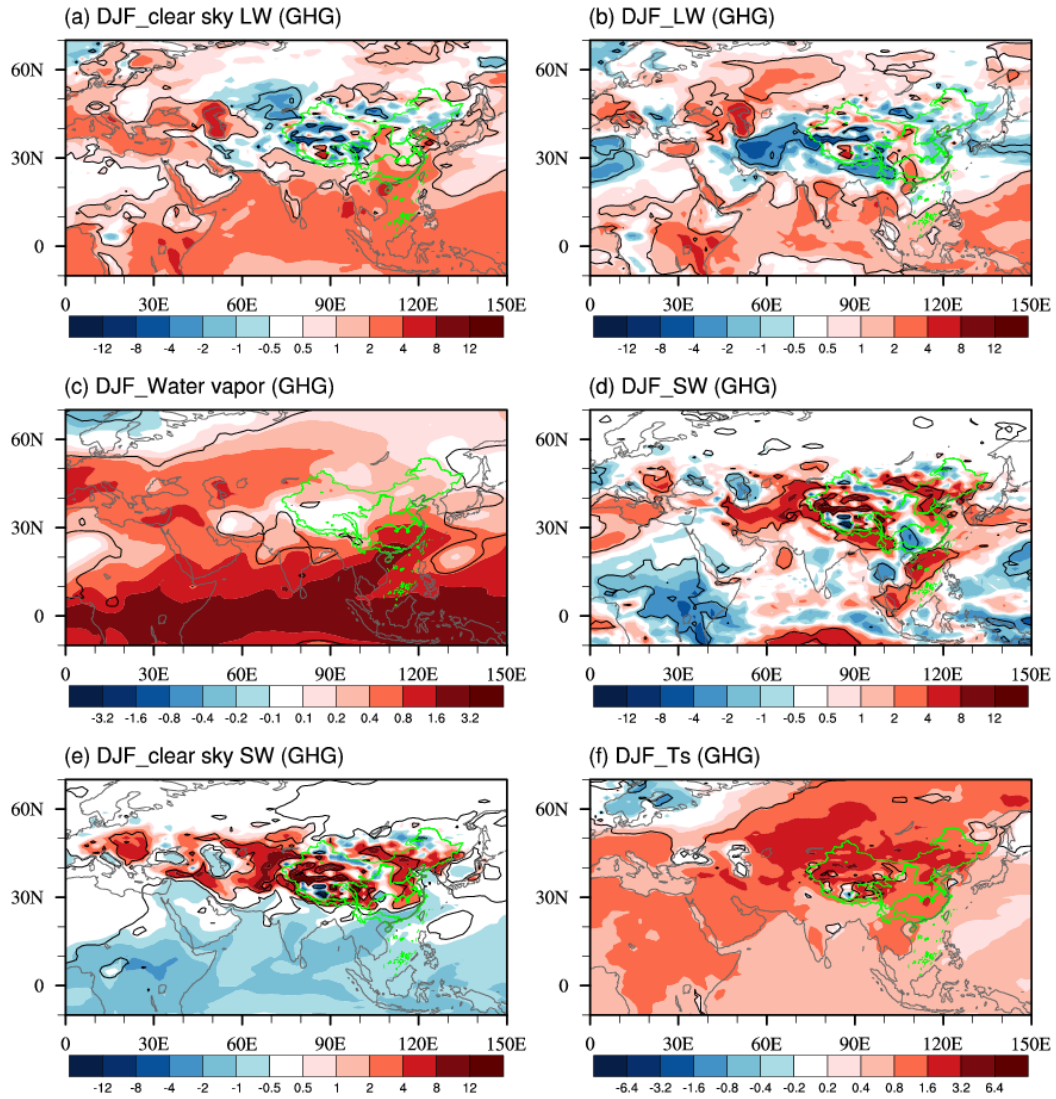




**Figure 7.** Spatial patterns of summer mean response to changes in GHG forcing: (a) clear sky LW radiation; (b) surface LW radiation; (c) total cloud cover (units: %) (d) low level cloud cover (units: %); (e) SW CRE; and (f) surface SW radiation. Radiation is the net component and in  $\text{W m}^{-2}$  with positive value meaning downward. The black lines highlight regions where the changes are statistically significant at the 90% confidence level base on *t*-test.

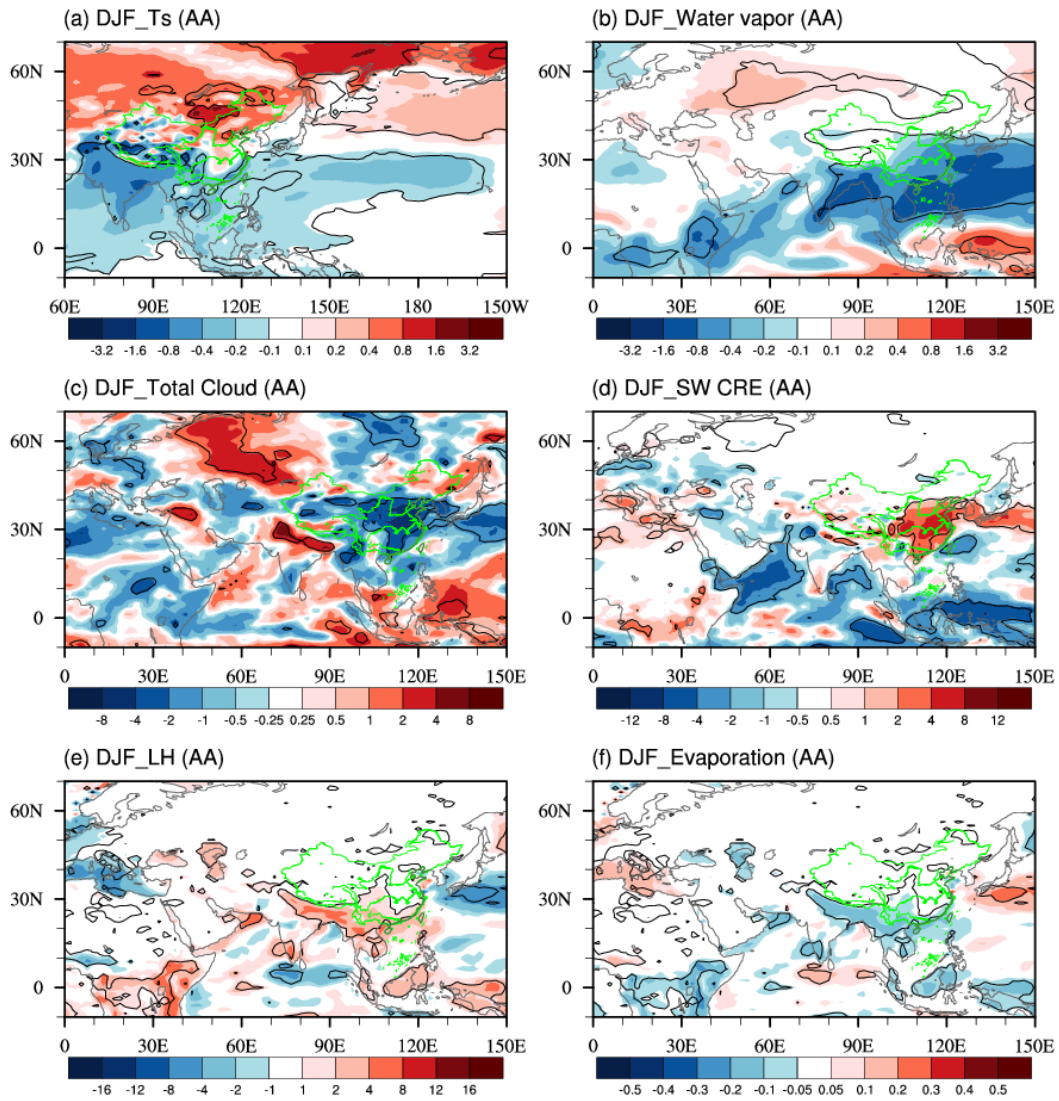


**Figure 8.** Spatial patterns of summer mean response to changes in AA forcing: (a) total AOD at 0.55  $\mu\text{m}$ ; (b) clear sky SW radiation; (c) SW CRE; (d) total cloud cover (units: %); (e) soil moisture (units:  $\text{kg m}^{-2}$ ); and (f) column-integrated water vapor (units:  $\text{kg m}^{-2}$ ). Radiation is the net component and in  $\text{W m}^{-2}$  with positive value meaning downward. The black lines highlight regions where the changes are statistically significant at the 90% confidence level base on  $t$ -test.



**Figure 9.** Spatial patterns of winter mean response to changes in GHG forcing: (a) clear sky LW radiation; (b) surface LW radiation; (c) column-integrated water vapor (units:  $kg m^{-2}$ ); (d) surface SW radiation; (e) clear sky SW radiation; and (f) skin temperature (units:  $^{\circ}C$ ). Radiation is the net component and in  $W m^{-2}$  with positive value meaning downward. The black lines highlight regions where the changes are statistically significant at the 90% confidence level base on *t*-test.





**Figure 10.** Spatial patterns of winter mean response to changes in AA forcing: (a) skin temperature (units:  $^{\circ}\text{C}$ ); (b) column-integrated water vapor (units:  $\text{kg m}^{-2}$ ); (c) total cloud cover (units: %); (d) SW CRE; (e) surface latent heat flux; and (f) evaporation (units:  $\text{kg m}^{-2}$ ). Radiation and flux are in  $\text{W m}^{-2}$  with positive value meaning downward. The black lines highlight regions where the changes are statistically significant at the 90% confidence level based on  $t$ -test.