

1 **Accelerated hydrological cycle over the Sanjiangyuan region induces**
2 **more streamflow extremes at different global warming levels**

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16 **Abstract.** Serving source water for the Yellow, Yangtze and Lancang-Mekong rivers,
17 the Sanjiangyuan region concerns 700 million people over its downstream areas.
18 Recent research suggests that the Sanjiangyuan region will become wetter in a
19 warming future, but future changes in streamflow extremes remain unclear due to the
20 complex hydrological processes over high-land areas and limited knowledge of the
21 influences of land cover change and CO₂ physiological forcing. Based on high
22 resolution land surface modeling during 1979~2100 driven by the climate and
23 ecological projections from 11 newly released Coupled Model Intercomparison
24 Project Phase 6 (CMIP6) climate models, we show that different accelerating rates of
25 precipitation and evapotranspiration at 1.5 °C global warming level induce 55% more
26 dry extremes over Yellow river and 138% more wet extremes over Yangtze river
27 headwaters compared with the reference period (1985~2014). An additional 0.5 °C
28 warming leads to a further nonlinear and more significant increase for both dry
29 extremes over Yellow river (22%) and wet extremes over Yangtze river (64%). The
30 combined role of CO₂ physiological forcing and vegetation greening, which used to
31 be neglected in hydrological projections, is found to alleviate dry extremes at 1.5 and
32 2.0 °C warming levels but to intensify dry extremes at 3.0 °C warming level. Moreover,
33 vegetation greening contributes half of the differences between 1.5 and 3.0 °C
34 warming levels. This study emphasizes the importance of ecological processes in
35 determining future changes in streamflow extremes, and suggests a “dry gets drier,
36 wet gets wetter” condition over headwaters.

37 **Keywords** Terrestrial hydrological cycle, streamflow extremes, global warming levels,

39 **1 Introduction**

40 Global temperature has increased at a rate of 1.7 °C/decade since 1970, contrary
41 to the cooling trend over the past 8000 years (Marcott et al., 2013). The temperature
42 measurements suggest that 2015-2019 is the warmest five years and 2010-2019 is also
43 the warmest decade since 1850 (WMO, 2020). To mitigate the impact of this
44 unprecedented warming on the global environment and human society, 195 nations
45 adopted the Paris Agreement which decides to “hold the increase in the global average
46 temperature to well below 2 °C above pre-industrial levels and pursuing efforts to limit
47 the temperature increase to 1.5 °C”.

48 The response of regional and global terrestrial hydrological processes, including
49 streamflow and its extremes, to different global warming levels has been investigated
50 by numerous studies in recent years (Chen et al., 2017; Döll et al., 2018; Marx et al.,
51 2018; Mohammed et al., 2017; Thober et al., 2018; Zhang et al., 2016). In addition to
52 climate change, recent works reveal the importance of the ecological factors (e.g., the
53 CO₂ physiological forcing and land cover change), which are often unaccounted for in
54 hydrological modeling works, in modulating the streamflow and its extremes. For
55 example, the increasing CO₂ concentration is found to alleviate the decreasing trend
56 of streamflow in the future at global scale, because the increased CO₂ concentration
57 will decrease the vegetation transpiration by reducing the stomatal conductance
58 (known as the CO₂ physiological forcing) (Fowler et al., 2019; Wiltshire et al., 2013;
59 Yang et al., 2019; Zhu et al., 2012). Contrary to the CO₂ physiological forcing, the
60 vegetation greening in a warming climate is found to have a significant role on

61 exacerbating hydrological drought, as it enhances transpiration and dries up the land
62 (Yuan et al., 2018b). However, the relative importance of CO₂ physiological forcing
63 and vegetation greening in influencing the terrestrial hydrology especially the
64 streamflow extremes is still unknown, and whether their combined impact changes at
65 different warming levels remains to be investigated.

66 Hosting the headwaters of the Yellow river, the Yangtze river and the
67 Lancang-Mekong river, the Sanjiangyuan region is known as the “Asian Water
68 Tower” and concerns 700 million people over its downstream. Changes of streamflow
69 and streamflow extremes over the Sanjiangyuan not only influence the local
70 ecosystem and water resource, but also affect the security of food, energy, and water
71 over the downstream areas. Both the regional climate and ecosystem show significant
72 changes over the Sanjiangyuan due to global warming (Bibi et al., 2018; Kuang and
73 Jiao, 2016; Liang et al., 2013; Yang et al., 2013; Zhu et al., 2016), which makes the
74 Sanjiangyuan a representative region to investigate the role of climate change and
75 ecological change (e.g., land use change and CO₂ physiological forcing) in
76 influencing the streamflow and its extremes (Cuo et al., 2014; Ji and Yuan, 2018; Zhu
77 et al., 2013). For example, historical changes in climate and ecology (e.g. land cover)
78 are found to cause significant reduction in mean and high flows during 1979-2005,
79 which potentially increases drought risk over its downstream areas (Ji and Yuan,
80 2018). And the CO₂ physiological forcing is revealed to cause equally large changes
81 in regional flood extremes as the precipitation over the Yangtze and Mekong rivers
82 (Fowler et al., 2019). Recent research suggests that the Sanjiangyuan region will

83 become warmer and wetter in the future, and extreme precipitation will also increase
84 at the 1.5 °C global warming level and further intensify with a 0.5 °C additional
85 warming (Li et al., 2018; Zhao et al., 2019). However, how the streamflow extremes
86 would respond to the 1.5 °C warming, what an additional 0.5 °C or even greater
87 warming would cause, and how much contributions do the ecological factors (e.g.,
88 CO₂ physiological forcing and land cover change) have, are still unknown. This
89 makes it difficult to assess the climate and ecological impact on this vital headwaters
90 region.

91 In this study, we investigate the future changes in the streamflow extremes over
92 the Sanjiangyuan region from an integrated eco-hydrological perspective by taking
93 CO₂ physiological forcing and land cover change into consideration. The combined
94 impacts of the above two ecological factors at different global warming levels are also
95 quantified and compared with the impact of climate change. The results will help
96 understand the role of ecological factors in future terrestrial hydrological changes
97 over the headwater regions like the Sanjiangyuan, and provide guidance and support
98 for the stakeholders to make relevant decisions and plans.

99 **2 Data and methods**

100 **2.1 Study domain and observational data**

101 The Sanjiangyuan region is located at the eastern part of the Tibetan Plateau
102 (Figure 1a), with the total area and mean elevation being 3.61×10^5 km² and 5000 m
103 respectively. It plays a critical role in providing fresh water, by contributing 35, 20
104 and 8% to the total annual streamflow of the Yellow, Yangtze and Lancang-Mekong

105 rivers (Li et al., 2017; Liang et al., 2013). The source regions of Yellow, Yangtze and
106 Lancang-Mekong rivers account for 46, 44 and 10% of the total area of the
107 Sanjiangyuan individually, and the Yellow river source region has a warmer climate
108 and sparser snow cover than the Yangtze river source region.

109 Monthly streamflow observations from the Tangnaihais (TNH) and the Zhimenda
110 (ZMD) hydrological stations (Figure 1a), which were provided by the local authorities,
111 were used to evaluate the streamflow simulations. Data periods are 1979-2011 and
112 1980-2008 for the Tangnaihais and Zhimenda stations individually. Monthly terrestrial
113 water storage change observation and its uncertainty during 2003-2014 was provided
114 by the Jet Propulsion Laboratory (JPL), which used the mass concentration blocks
115 (mascons) basis functions to fit the Gravity Recovery and Climate Experiment
116 (GRACE) satellite's inter-satellite ranging observations (Watkins et al., 2015). The
117 Model Tree Ensemble evapotranspiration (MTE_ET; Jung et al., 2009) and the Global
118 Land Evaporation Amsterdam Model evapotranspiration (GLEAM_ET) version 3.3a
119 (Martens et al., 2017) were also used to evaluate the model performance on ET
120 simulation.

121 **2.2 CMIP6 Data**

122 Here, 19 Coupled Model Intercomparison Project phase 6 (CMIP6, Eyring et al.,
123 2016) models which provide precipitation, near-surface temperature, specific
124 humidity, 10-m wind speed, surface downward shortwave and longwave radiations at
125 daily timescale were first selected for evaluation. Then, models were chosen for the
126 analysis when the simulated meteorological forcings (e.g., precipitation, temperature,

127 humidity, and shortwave radiation) averaged over the Sanjiangyuan region have the
128 same trend sign as the observation during 1979-2014. Table 1 shows the 11 CMIP6
129 models that were finally chosen in this study. For the future projection (2015-2100),
130 we chose two Shared Socioeconomic Pathways (SSP) experiments: SSP585 and
131 SSP245. SSP585 combines the fossil-fueled development socioeconomic pathway
132 and 8.5W/m^2 forcing pathway (RCP8.5), while SSP245 combines the moderate
133 development socioeconomic pathway and 4.5 W/m^2 forcing pathway (RCP4.5)
134 (O'Neill et al., 2016). Land cover change is quantified by leaf area index (LAI) as
135 there is no significant transition between different vegetation types (not shown)
136 according to the Land-use Harmonization 2 (LUH2) dataset
137 (<https://esgf-node.llnl.gov/search/input4mips/>). For the CNRM-CM6-1, FGOALS-g3
138 and CESM2, the ensemble mean of LAI simulations from the other 8 CMIP6 models
139 was used because CNRM-CM6-1 and FGOALS-g3 do not provide dynamic LAI
140 while the CESM2 simulates an abnormally large LAI over the Sanjiangyuan region.
141 To avoid systematic bias in meteorological forcing, the trend-preserved bias
142 correction method suggested by ISI-MIP (Hempel et al., 2013), was applied to the
143 CMIP6 model simulations at monthly scale. The China Meteorological Forcing
144 Dataset (CMFD; He et al., 2020) is taken as meteorological observation. For each
145 month, temperature bias in CMIP6 simulations during 1979-2014 was directly
146 deducted. Future temperature simulations in SSP245 and SSP585 experiments were
147 also adjusted according to the historical bias. Other variables were corrected by using
148 a multiplicative factor, which was calculated by using observations to divide

149 simulation during 1979-2014. In addition, monthly leaf area index was also adjusted
150 to be consistent with satellite observation using the same method as temperature. All
151 variables were first interpolated to the 10 km resolution over the Sanjiangyuan region
152 and the bias correction was performed for each CMIP6 model at each grid. After bias
153 correction, absolute changes of temperature and leaf area index, and relative changes
154 of other variables were preserved at monthly time scale (Hempel et al., 2013). Then,
155 the adjusted CMIP6 daily meteorological forcings were disaggregated into hourly
156 using the diurnal cycle ratios from the China Meteorological Forcing Dataset.

157 The historical CO₂ concentration used here is the same as the CMIP6 historical
158 experiment (Meinshausen et al., 2017), while future CO₂ concentration in SSP245 and
159 SSP585 scenarios came from simulations of a reduced-complexity carbon-cycle
160 model MAGICC7.0 (Meinshausen et al., 2020).

161 **2.3 Experimental design**

162 The land surface model used in this study is the Conjunctive Surface-Subsurface
163 Process model version 2 (CSSPv2), which has been proved to simulate the energy and
164 water processes over the Sanjiangyuan region well (Yuan et al., 2018a). Figure 2
165 shows the structure and main ecohydrological processes in CSSPv2. The CSSPv2 is
166 rooted in the Common Land Model (CoLM; Dai et al., 2003) with some
167 improvements at hydrological processes. CSSPv2 has a volume-averaged soil
168 moisture transport (VAST) model, which solves the quasi-three dimensional
169 transportation of the soil water and explicitly considers the variability of moisture flux
170 due to subgrid topographic variations (Choi et al., 2007). Moreover, the Variable

171 Infiltration Capacity runoff scheme (VIC; Liang et al., 1994), and the influences of
 172 soil organic matters on soil hydrological properties are incorporated into the CSSPv2
 173 by Yuan et al. (2018a), to improve its performance in simulating the terrestrial
 174 hydrology over the Sanjiangyuan region. Similar to the CoLM and the Community
 175 Land Surface Model (CLM; Oleson et al., 2013), vegetation transpiration in CSSPv2
 176 is based on Monin-Obukhov similarity theory, and the transpiration rate is constrained
 177 by leaf boundary layer and stomatal conductances. Parameterization of the stomatal
 178 conductance (g_s) in CSSPv2 is

$$179 \quad g_s = m \frac{A_n}{P_{CO_2} / P_{am}} h_s + b \beta_t$$

180 where the m is a plant functional type dependent parameter, A_n is leaf net
 181 photosynthesis ($\mu mol CO_2 m^{-2} s^{-1}$), P_{CO_2} is the CO_2 partial pressure at the leaf
 182 surface (Pa), P_{am} is the atmospheric pressure (Pa), h_s is the leaf surface
 183 humidity, b is the minimum stomatal conductance ($\mu mol m^{-2} s^{-1}$), while β_t is the
 184 soil water stress function. Generally, the stomatal conductance decreases with the
 185 increasing of CO_2 concentration.

186 First, bias-corrected meteorological forcings from CMIP6 historical experiment
 187 were used to drive the CSSPv2 model (CMIP6_His/CSSPv2). All simulations were
 188 conducted for two cycles during 1979-2014 at half-hourly time step and 10 km spatial
 189 resolution, with the first cycle serving as the spin-up. Correlation coefficient (CC) and
 190 root mean squared error (RMSE) were calculated for the observed and simulated
 191 monthly streamflow, annual evapotranspiration and monthly terrestrial water storage,
 192 to evaluate the model performance. The King-Gupta efficiency (KGE; Gupta et al.,

193 2009), which is widely used in streamflow evaluations, was also calculated for
 194 streamflow simulations. Above metrics were calculated as follows:

$$195 \quad CC = \frac{\sum_{i=1}^n (x_i - \bar{x})(y_i - \bar{y})}{\sqrt{\sum_{i=1}^n (x_i - \bar{x})^2 \sum_{i=1}^n (y_i - \bar{y})^2}}$$

$$196 \quad RMSE = \sqrt{\frac{\sum_{i=1}^n (x_i - y_i)^2}{n}}$$

$$197 \quad KGE = 1 - \sqrt{(1 - CC)^2 + (1 - \frac{\sigma_x}{\sigma_y})^2 + (1 - \frac{\bar{x}}{\bar{y}})^2}$$

198 where x_i and y_i are observed and simulated variables in a specific month/year i
 199 individually, and \bar{x} and \bar{y} are corresponding monthly/annual means during the
 200 whole evaluation period n . The σ_x and σ_y are observed and simulated standard
 201 deviations respectively. The correlation coefficient represents the correlation between
 202 simulation and observation, while RMSE means simulated error. The KGE ranges
 203 from negative infinity to 1 and model simulations can be regard as satisfactory when
 204 the KGE is larger than 0.5 (Moriasi et al., 2007).

205 Second, bias-corrected meteorological forcings in SSP245 and SSP585 were
 206 used to drive CSSPv2 during 2015-2100 with dynamic LAI and CO₂ concentration
 207 (CMIP6_SSP/CSSPv2). Initial conditions of CMIP6_SSP/CSSPv2 came from the last
 208 year in CMIP6_His/CSSPv2.

209 Then, the second step was repeated twice by fixing the monthly LAI
 210 (CMIP6_SSP/CSSPv2_FixLAI) and mean CO₂ concentration
 211 (CMIP6_SSP/CSSPv2_FixCO₂) at 2014 level. The difference between

212 CMIP6_SSP/CSSPv2 and CMIP6_SSP/CSSPv2_FixLAI is regarded as the net effect
213 of land cover change, and the difference between CMIP6_SSP/CSSPv2 and
214 CMIP6_SSP/CSSPv2_FixCO2 is regarded as the net effect of CO₂ physiological
215 forcing.

216 **2.4 Warming level determination**

217 A widely used time-sampling method was adopted to determine the periods of
218 different global warming levels (Chen et al., 2017; Döll et al., 2018; Marx et al., 2018;
219 Mohammed et al., 2017; Thober et al., 2018). According to the HadCRUT4 dataset
220 (Morice et al., 2012), the global mean surface temperature has increased by 0.66 °C
221 from the pre-industrial era (1850-1900) to the reference period defined as 1985-2014.
222 Then, starting from 2015, 30-years running mean global temperatures were compared
223 to those of the 1985-2014 period for each GCM simulation. And the
224 1.5 °C/2.0 °C/3.0 °C warming period is defined as the 30-years period when the
225 0.84 °C/1.34 °C/2.34 °C global warming, compared with the reference period
226 (1985-2014), is first reached. The median years of identified 30-year periods, referred
227 as “crossing years”, are shown in Table 2.

228 **2.5 Definition of dry and wet extremes and robustness assessment**

229 In this research, the standardized streamflow index (SSI) was used to define dry
230 and wet extremes (Vicente-Serrano et al., 2012; Yuan et al., 2017). A gamma
231 distribution was first fitted using July-September (flood season) mean streamflow
232 during the reference period. Then the fitted distribution was used to calculate the
233 standardized deviation of the July-September mean streamflow (i.e. SSI) in each year

234 during both the reference and projection periods. Here, dry and wet extremes were
235 defined as where SSIs are smaller than -1.28 (a probability of 10%) and larger than
236 1.28 respectively.

237 The relative changes of dry/wet extremes frequencies between the reference
238 period and different warming periods were first calculated for each GCMs under each
239 SSP scenarios, and the ensemble means were then determined for each warming
240 levels. To quantify the uncertainty, above calculations were repeated by doing
241 bootstrapping 10,000 times, and 11 GCMs were resampled with replacement during
242 each bootstrap (Christopher et al., 2018). The 5% and 95% percentiles of the total
243 10,000 estimations were finally taken as the 5~95% uncertainty ranges.

244 **3 Results**

245 **3.1 Terrestrial hydrological changes at different warming levels**

246 As shown in Figures 1b-1e, observations (pink lines) show that the annual
247 temperature, precipitation and growing season LAI increase at the rates of
248 0.63 °C/decade ($p=0$), 16.9 mm/decade ($p=0.02$), and 0.02 m²/m²/decade ($p=0.001$)
249 during 1979-2014 respectively. The ensemble means of CMIP6 simulations (black
250 lines) can generally capture the historical increasing trends of temperature
251 (0.30 °C/decade, $p=0$), precipitation (7.1 mm/decade, $p=0$) and growing season LAI
252 (0.029 m²/m²/decade, $p=0$), although the increasing trends of precipitation and
253 temperature are underestimated. In 2015-2100, the SSP245 scenario (blue lines)
254 shows continued warming, wetting and greening trends, and the trends are larger in
255 the SSP585 scenario (red lines). The CO₂ concentration also keeps increasing during

256 2015-2100 and reaches to 600 ppm and 1150 ppm in 2100 for the SSP245 and
257 SSP585 scenarios respectively. Although the SSP585 scenario reaches the same
258 warming levels earlier than the SSP245 scenario (Table 2), there is no significant
259 difference between them in the meteorological variables during the same warming
260 period (not shown). Thus, we do not distinguish SSP245 and SSP585 scenarios at the
261 same warming level in the following analysis.

262 Figure 3 and Table 3 show the evaluation of model simulation. Driven by
263 observed meteorological and ecological forcings, the CMFD/CSSPv2 simulates
264 monthly streamflow over the Yellow and Yangtze river headwaters quite well.
265 Compared with the observation at Tangnaihai (TNH) and Zhimenda (ZMD) stations,
266 the Kling-Gupta efficiencies of the CMFD/CSSPv2 simulated monthly streamflow are
267 0.94 and 0.91 respectively. The simulated monthly Terrestrial Water Storage Anomaly
268 (TWSA) during 2003-2014 in CMFD/CSSPv2 also agrees with the GRACE satellite
269 observation and captures the increasing trend. For the interannual variations of
270 evapotranspiration, CMFD/CSSPv2 is consistent with the ensemble mean of the
271 GLEAM_ET and MTE_ET products, and the correlation coefficient and root mean
272 squared error (RMSE) during 1982-2011 are 0.87 ($p < 0.01$) and 14 mm/year
273 respectively. This suggests the good performance of the CSSPv2 in simulating the
274 hydrological processes over the Sanjiangyuan region. Although meteorological and
275 ecological outputs from CMIP6 models have coarse resolutions (~100km), the land
276 surface simulation driven by bias corrected CMIP6 results (CMIP6_His/CSSPv2) also
277 captures the terrestrial hydrological variations reasonably well. The Kling-Gupta

278 efficiency of the ensemble mean streamflow simulation reaches up to 0.71~0.81, and
279 the ensemble mean monthly Terrestrial Water Storage Anomaly (TWSA) and annual
280 evapotranspiration generally agree with observations and other reference data
281 (Figures 3c-d).

282 Figure 4 shows relative changes of terrestrial hydrological variables over the
283 Sanjiangyuan region at different warming levels. The ensemble mean of the increase
284 in annual precipitation is 5% at 1.5 °C warming level, and additional 0.5 °C and 1.5 °C
285 warming will further increase the wetting trends to 7% and 13% respectively. Annual
286 evapotranspiration experiences significant increases at all warming levels, and the
287 ensemble mean increases are 4%, 7% and 13% at 1.5, 2.0 and 3.0 °C warming levels
288 respectively. The ratio of transpiration to evapotranspiration also increases
289 significantly, indicating that vegetation transpiration increases much larger than the
290 soil evaporation and canopy evaporation. Although annual total runoff has larger
291 relative changes than evapotranspiration (6%, 9% and 14% at 1.5, 2.0 and 3.0 °C
292 warming levels respectively), the uncertainty is large as only 75% of the models show
293 positive signals, which may be caused by large uncertainty in the changes during
294 summer and autumn seasons. The terrestrial water storage (TWS) which includes
295 foliage water, surface water, soil moisture and groundwater, shows slightly decreasing
296 trend at annual scale, suggesting that the increasing precipitation in the future
297 becomes extra evapotranspiration and runoff instead of recharging the local water
298 storage. The accelerated terrestrial hydrological cycle also exists at seasonal scale, as
299 the seasonal changes are consistent with the annual ones.

300 **3.2 Changes in streamflow extremes at different warming levels**

301 Although the intensified terrestrial hydrology induces more streamflow over the
302 headwater region of Yellow river during winter and spring months, streamflow does
303 not increase and even decreases during the flood season (July-September; Figure 5a).
304 Figure 5b shows the changes of streamflow dry extremes over the Yellow river source
305 region at different warming levels, with the error bars showing estimated uncertainties.
306 The frequency of streamflow dry extremes over the Yellow river is found to increase
307 by 55% at 1.5 °C warming level (Figure 5b), but the uncertainty is larger than the
308 ensemble mean. However, the dry extreme frequency will further increase to 77% and
309 125% at the 2.0 and 3.0 °C warming levels and the results become significant (Figure
310 5b). No significant changes are found for the wet extremes at all warming levels over
311 the Yellow River headwater region, as the uncertainty ranges are larger than the
312 ensemble means.

313 Over the Yangtze river headwater region, streamflow increases in all months at
314 different warming levels (Figure 5c). The frequency of wet extremes increases
315 significantly by 138%, 202% and 232% at 1.5, 2.0 and 3.0 °C warming levels (Figure
316 4d), suggesting a higher risk of flooding. Although the frequency of dry extremes
317 tends to decrease significantly by 35%, 44%, 34% at the three warming levels, the
318 changes are much smaller than those of the wet extremes. Moreover, contributions
319 from climate change and ecological change are both larger than the uncertainty ranges
320 (not shown), suggesting that their impacts on the changes of dry extremes over the
321 Yangtze river headwater region are not distinguishable. Thus, we mainly focus on the

322 dry extremes over the Yellow river and the wet extremes over the Yangtze river in the
323 following analysis.

324 Different changes of streamflow extremes over the Yellow and Yangtze rivers
325 can be interpreted from different accelerating rates of precipitation and
326 evapotranspiration. Figure 6 shows probability density functions (PDFs) of
327 precipitation, evapotranspiration and their difference (P-ET, i.e. residual water for
328 runoff generation) during the flood season. Over the Yellow river, PDFs of
329 precipitation and evapotranspiration both shift to the right against the reference period,
330 except for the precipitation at 1.5 °C warming level. However, the increasing trend of
331 evapotranspiration is stronger than that of precipitation, leading to a left shift of PDF
332 for P-ET. Moreover, increased variations of precipitation and evapotranspiration, as
333 indicated by the increased spread of their PDFs, also lead to a larger spread of PDFs
334 of P-ET. The above two factors together induce a heavier left tail in the PDF of P-ET
335 for the warming future than the reference period (Figure 6e). The probability of
336 $P-ET < 80\text{mm}$ increases from 0.1 during historical period to 0.11, 0.13 and 0.16 at 1.5,
337 2.0 and 3.0 °C warming levels individually. This indicates a higher probability of less
338 water left for runoff generation at different warming levels, given little changes in
339 TWS (section 3.1). Moreover, Figure 6e also shows little change to the right tails in
340 the PDF of P-ET as probability for $P-ET > 130\text{mm}$ stays around 0.1 at different
341 warming levels, suggesting little change to the probability of high residual water. This
342 is consistent with the insignificant wet extreme change over the Yellow river. Over the
343 Yangtze river, however, intensified precipitation is much larger than the increased

344 evapotranspiration, leading to a systematic rightward shift of the PDF of P-ET
345 (Figures 6b, 6d and 6f). Thus both the dry and wet extremes show significant changes
346 over the Yangtze river.

347 **3.3 Influences of land cover change and CO₂ physiological forcing**

348 Figures 7a-7b show the changes of streamflow extremes (compared with the
349 reference period) induced by climate and ecological factors. Although the contribution
350 from climate change (red bars in Figures. 7a-7b) is greater than the ecological factors
351 (blue and cyan bars in in Figures. 7a-7b), influences of CO₂ physiological forcing and
352 land cover change are nontrivial. The CO₂ physiological forcing tends to alleviate dry
353 extremes (or increase wet extremes), while land cover change plays a contrary role.
354 Over the Yellow river, the combined impact of the two ecological factors (sum of blue
355 and cyan bars) reduces the increasing trend of dry extremes caused by climate change
356 (red bars) by 18~22% at 1.5 and 2.0 °C warming levels, while intensifies the dry
357 extremes by 9% at 3.0 °C warming level. This can be interpreted from their
358 contributions to the evapotranspiration, as the increased LAI enhancement on ET is
359 weaker than the suppression effect of CO₂ physiological impact at 1.5 and 2.0 °C
360 warming levels, while stronger at 3.0 °C warming level (not shown). Over the Yangtze
361 river, similarly, combined effect of land cover and CO₂ physiological forcing
362 increases the wet extremes by 9% at 1.5 °C warming level while decreases the wet
363 extremes by 12% at 3.0 °C warming level.

364 In addition, Figures 7c and 7d show that the combined impact of CO₂
365 physiological forcing and land cover change also influences the differences between

366 different warming levels. Over the Yellow river, climate change increases dry
367 extremes by 26% from 1.5 to 2.0 °C warming level, and by 40% from 1.5 and 3.0 °C
368 warming level (red bars in Figure 7c). After considering the two ecological factors
369 (pink bars in Figure 7c), above two values change to 22% and 70% respectively, and
370 the difference between 1.5 and 3.0 °C warming levels becomes significant. For the wet
371 extreme over the Yangtze river (Figure 7d), the climate change induced difference
372 between 1.5 and 2.0 °C warming levels is decreased by 16% after accounting for the
373 two ecological factors. And this decrease reaches up to 49% for the difference
374 between 1.5 and 3.0 °C warming levels. We also compared the scenarios when CO₂
375 physiological forcing and land cover change are combined with climate change
376 individually (blue and cyan bars in Figures 7c-d), and the results show the land cover
377 change dominates their combined influences on the difference between different
378 warming levels.

379 **4 Conclusions and Discussion**

380 This study investigates changes of streamflow extremes over the Sanjiangyuan
381 region at different global warming levels through high-resolution land surface
382 modeling driven by CMIP6 climate simulations. The terrestrial hydrological cycle
383 under global warming of 1.5 °C is found to accelerate by 4~6% compared with the
384 reference period of 1985-2014, according to the relative changes of precipitation,
385 evapotranspiration and total runoff. The terrestrial water storage, however, shows
386 slight but significant decreasing trend as increased evapotranspiration and runoff are
387 larger than the increased precipitation. This decreasing trend of terrestrial water

388 storage in the warming future is also found in six major basins in China (Jia et al.,
389 2020). Although streamflow changes during the flood season has a large uncertainty,
390 the frequency of wet extremes over the Yangtze river will increase significantly by
391 138% and that of dry extremes over the Yellow river will increase by 55% compared
392 with that during 1985~2014. With an additional 0.5 °C warming, the frequency of dry
393 and wet extremes will increase further by 22~64%. If the global warming is not
394 adequately managed (e.g., to reach 3.0 °C), wet extremes over the Yangtze river and
395 dry extremes over the Yellow river will increase by 232% and 125%. The changes
396 from 1.5 to 2.0 and 3.0 °C are nonlinear compared with that from reference period to
397 1.5 °C, which are also found for some fixed-threshold climate indices over the Europe
398 (Dosio and Fischer, 2018). It is necessary to cap the global warming at 2 °C or even
399 lower level, to reduce the risk of wet and dry extremes over the Yangtze and Yellow
400 rivers.

401 This study also shows the nontrivial contributions from land cover change and
402 CO₂ physiological forcing to the extreme streamflow changes especially at 2.0 and
403 3.0 °C warming levels. The CO₂ physiological forcing is found to increase streamflow
404 and reduce the dry extreme frequency by 14~24%, which is consistent with previous
405 research that CO₂ physiological forcing would increase available water and reduce
406 water stress at the end of this century (Wiltshire et al., 2013). However, our results
407 further show that the drying effect of increasing LAI on streamflow will exceed the
408 wetting effect of CO₂ physiological forcing at 3.0 °C warming level (during
409 2048~2075) over the Sanjiangyuan region, making a reversion in the combined

410 impacts of CO₂ physiological forcing and land cover. Thus it is vital to consider the
411 impact of land cover change in the projection of future water stress especially at high
412 warming scenarios.

413 Moreover, about 43~52% of the extreme streamflow changes between 1.5 and
414 3.0 °C warming levels are attributed to the increased LAI. Considering the LAI
415 projections from different CMIP6 models are induced by the climate change, it can be
416 inferred that the indirect influence of climate change (e.g., through land cover change)
417 has the same and even larger importance on the changes of streamflow extremes
418 between 1.5 and 3.0 °C or even higher warming levels, compared with the direct
419 influence (e.g., through precipitation and evapotranspiration). Thus, it is vital to
420 investigate hydrological and its extremes changes among different warming levels
421 from an eco-hydrological perspective instead of focusing on climate change alone.

422 Although we used 11 CMIP6 models combined with two SSP scenarios to reduce
423 the uncertainty of future projections caused by GCMs, using a single land surface
424 model may lead to some uncertainties (Marx et al., 2018). However, considering the
425 high performance of the CSSPv2 land surface model over the Sanjiangyuan region
426 and the dominate role of GCMs' uncertainty over this region (Zhao et al., 2019;
427 Samaniego et al., 2017), uncertainty from the CSSPv2 model should not influence the
428 robust of the result.

429

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435

436 **Competing interests**

437 The authors declare that they have no conflict of interest.

438

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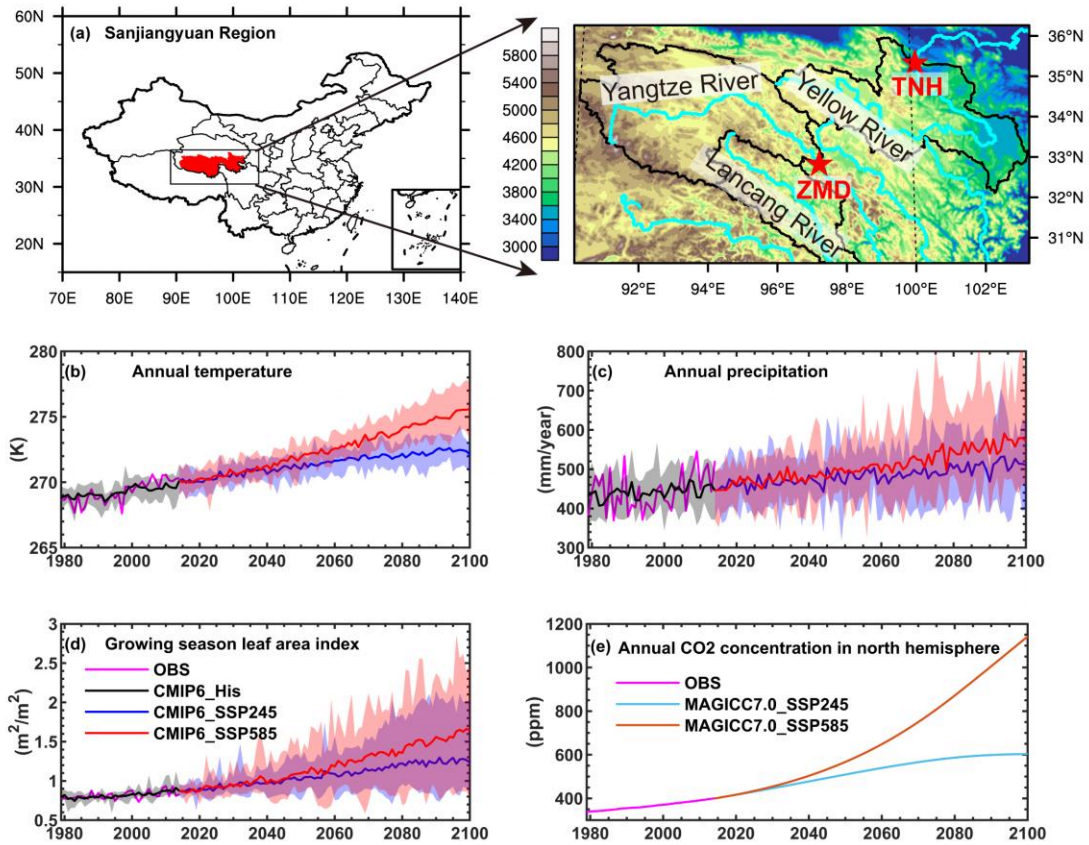
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609



610

611 **Figure 1.** (a) The locations of the Sanjiangyuan region and streamflow gauges. (b)-(e)

612 are the time series of annual temperature, precipitation, growing season leaf area

613 index and CO₂ concentration averaged over the Sanjiangyuan region during

614 1979-2100. Red pentagrams in (a) are two streamflow stations named Tangnaihai

615 (TNH) and Zhimenda (ZMD). Black, blue and red lines in (b-d) are ensemble means

616 of CMIP6 model simulations from the historical, SSP245 and SSP585 experiments.

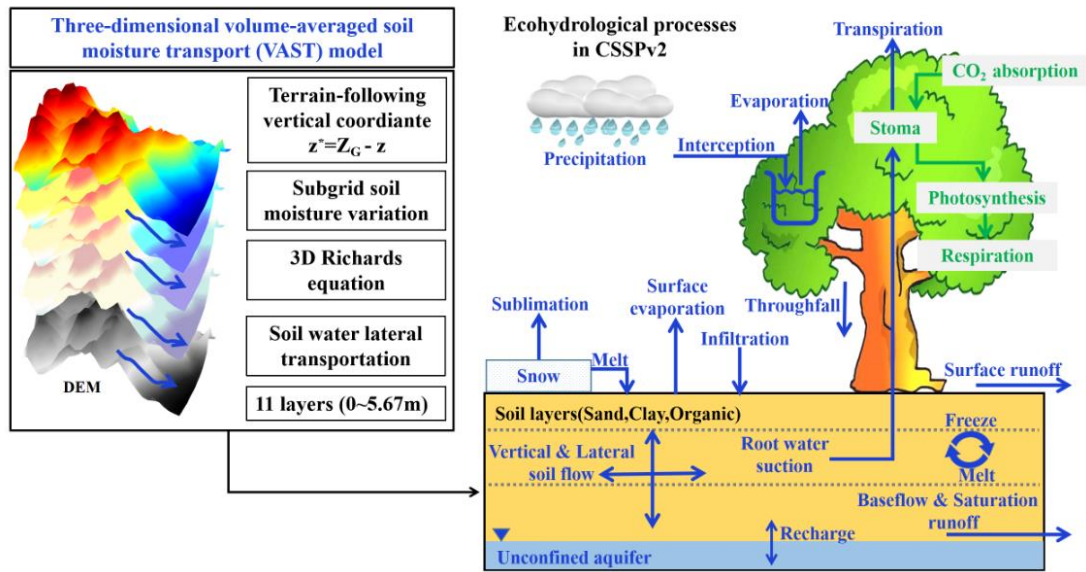
617 Shadings are ranges of individual ensemble members. Cyan and brown lines in (e) are

618 future CO₂ concentration under SSP245 and SSP585 scenarios simulated by

619 MAGICC7.0 model.

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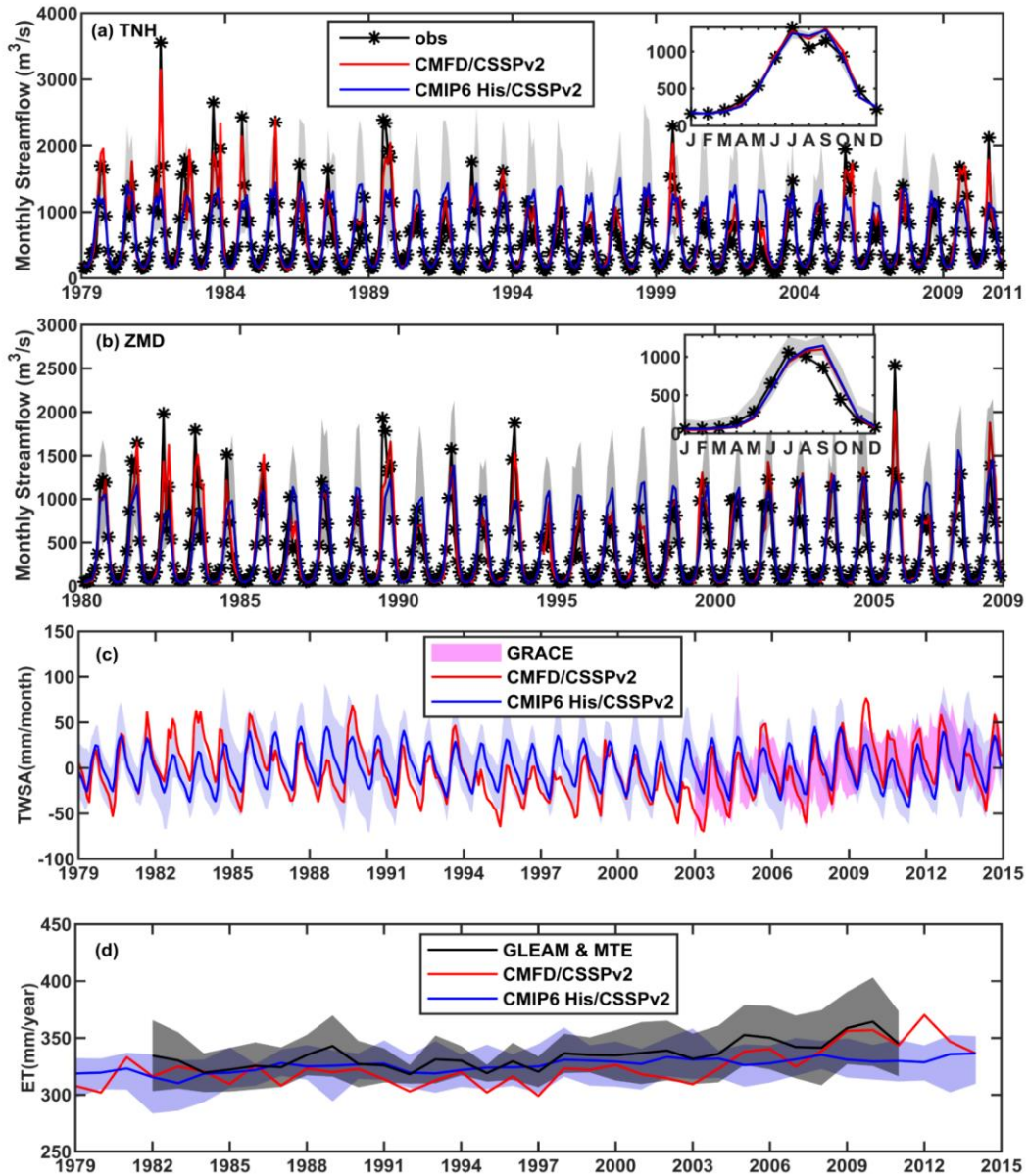
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622

623 **Figure 2.** Structure and main ecohydrological processes in the Conjointive

624 Surface-Subsurface Process version 2 (CSSPv2) land surface model.



625

626 **Figure 3.** Evaluation of model simulations. (a-b) Observed and simulated monthly

627 streamflow at the Tangnaihai (TNH) and Zhimenda (ZMD) hydrological stations, with

628 the climatology shown in the upper-right corner. (c-d) Evaluation of the simulated

629 monthly terrestrial water storage anomaly (TWSA) and annual evapotranspiration (ET)

630 averaged over the Sanjiangyuan region. Red lines are CSSPv2 simulation forced by

631 observed meteorological forcing. Blue lines represent ensemble means of 11

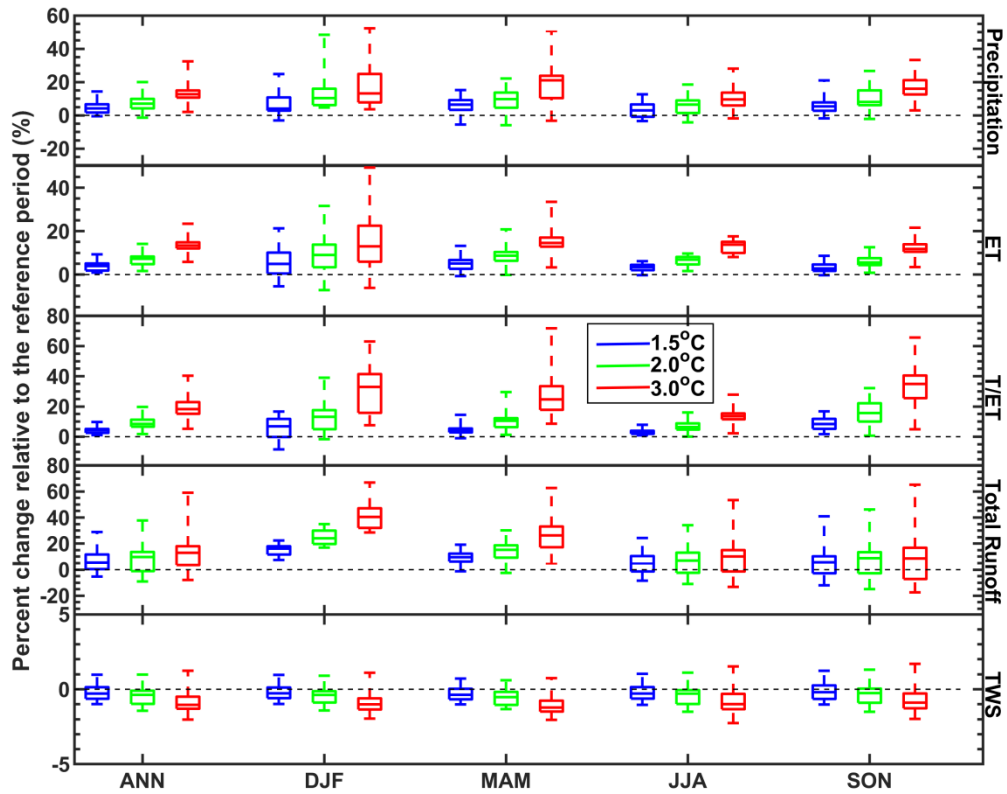
632 CMIP6_His/CSSPv2 simulations, while gray shadings in (a-b) and blue shadings in

633 (c-d) are ranges of individual ensemble members. Pink shading in (c) is GRACE

634 satellite observations. Black line and black shading in (d) are ensemble mean and
635 ranges of GLEAM_ET and MTE_ET datasets.

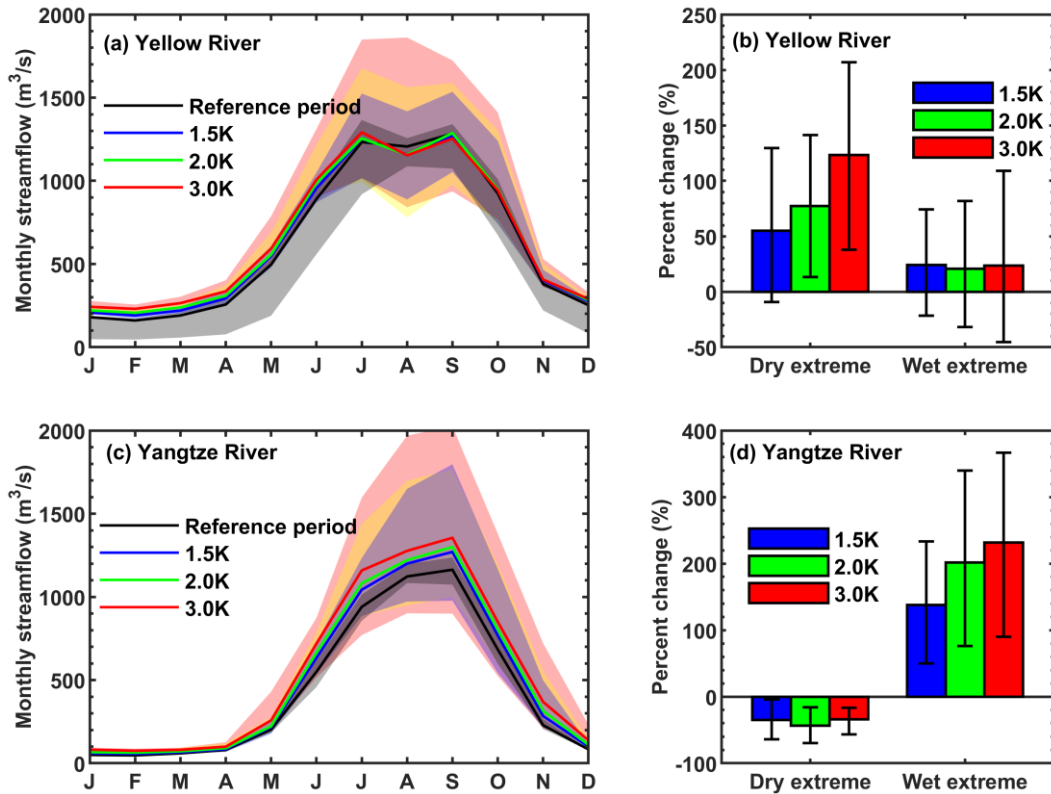
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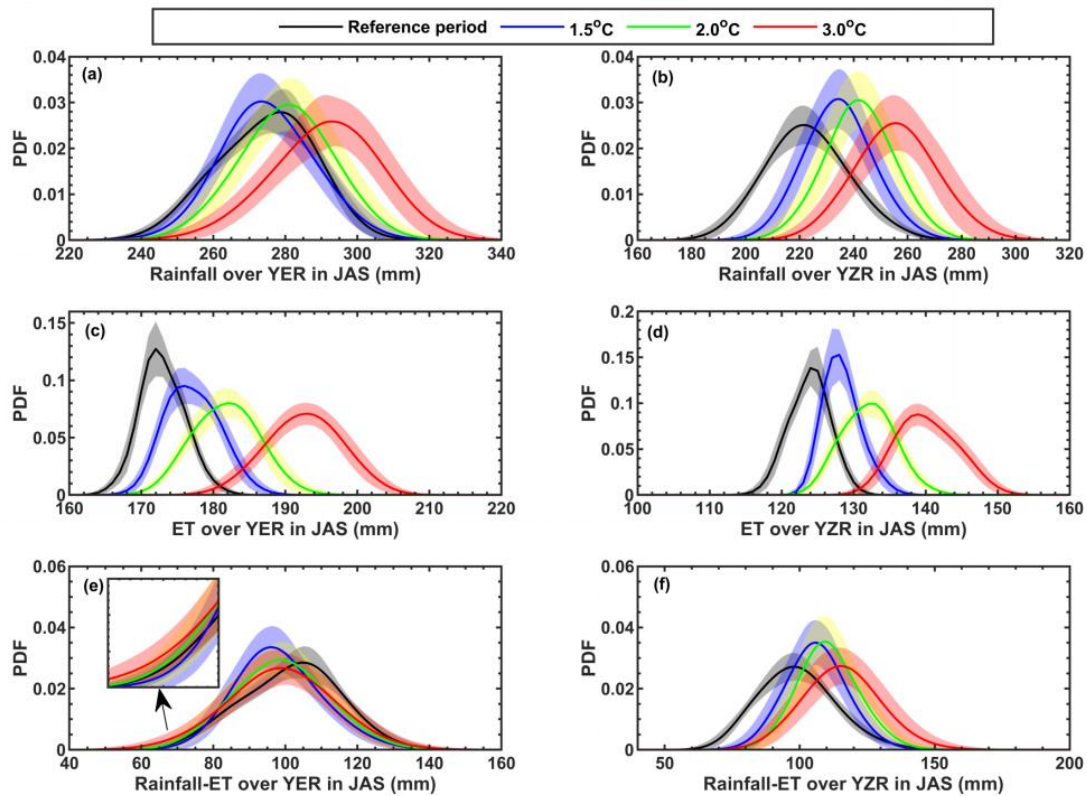
638

639 **Figure 4.** Box plots of relative changes of regional mean precipitation,
 640 evapotranspiration (ET), ratio of transpiration to evapotranspiration (T/ET), total
 641 runoff and terrestrial water storage (TWS) at different global warming levels.
 642 Reference period is 1985-2014, and annual (ANN) and seasonal (winter: DF, spring:
 643 MAM, summer: JJA and autumn: SON) results are all shown. Boxes show 25th to
 644 75th ranges among 22 CMIP6_SSP/CSSPv2 simulations, while lines in the boxes are
 645 median values.



647

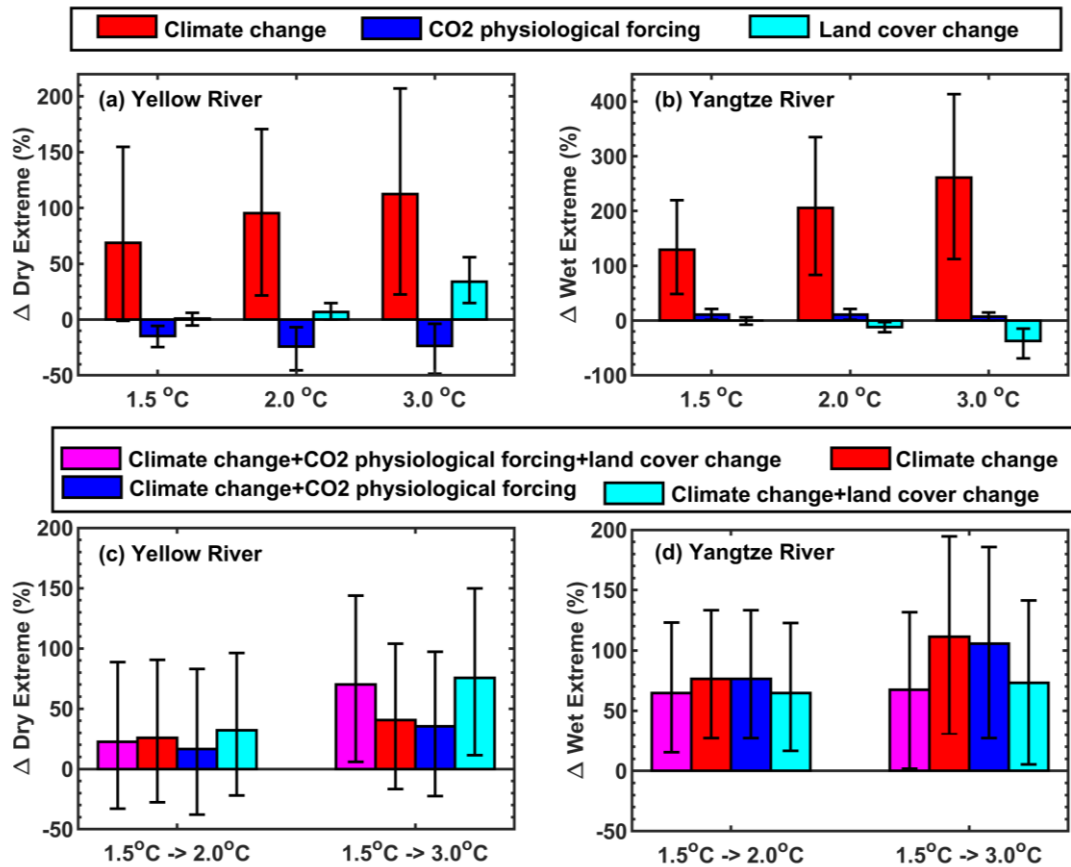
648 **Figure 5.** Changes of streamflow and its extremes at the outlets of the headwater
 649 regions of the Yellow river and the Yangtze river, i.e., Tangnaihai gauge and
 650 Zhimenda gauge. (a) Simulated monthly streamflow over the Yellow river during the
 651 reference period (1985-2014) and the periods with different global warming levels.
 652 Solid lines represent ensemble means, while shadings are ranges of individual
 653 ensemble members. (b) Percent changes in frequency of dry and wet extremes in
 654 July-September at different warming levels. Colored bars are ensemble means, while
 655 error bars are 5~95% uncertainty ranges estimated by using bootstrapping for 10,000
 656 times. (c) and (d) are the same as (a) and (b), but for the Yangtze river.



658

659 **Figure 6.** Probability density functions (PDFs) of regional mean rainfall,
 660 evapotranspiration (ET) and their difference over the headwater regions of Yellow
 661 river (YER) and Yangtze river (YZR) during flooding seasons (July-September) for
 662 the reference period (1985-2014) and the periods with 1.5, 2.0 and 3.0 °C global
 663 warming levels. Shadings are 5~95% uncertainty ranges.

664



665

666 **Figure 7.** (a-b) Influences of climate change, CO₂ physiological forcing and land
 667 cover change on relative changes in frequency of the dry and wet extremes in
 668 July-September at different global warming levels for the headwater regions of
 669 Yellow river and Yangtze river. (c-d) Changes of dry and wet extremes under
 670 additional warming of 0.5 °C and 1.5 °C with the consideration of different factors. All
 671 the changes are relative to the reference period (1985-2014). Ensemble means are
 672 shown by colored bars while the 5~95% uncertainty ranges estimated by using
 673 bootstrapping for 10,000 times are represented by error bars.

674

675 **Table 1.** CMIP6 simulations used in this study. His means historical simulations
676 during 1979-2014 with both anthropogenic and natural forcings, SSP245 and SSP585
677 represent two Shared Socioeconomic Pathways during 2015-2100. Note the
678 CNRM-CM6-1 and CNRM-ESM2-1 do not provide r1i1p1f1 realization, so r1i1p1f2
679 was used instead.

No.	Models	Experiments	Realization	Horizontal Resolution (Longitude × Latitude Grid Points)
1	ACCESS-ESM1-5	His/SSP245/SSP585	r1i1p1f1	192×145
2	BCC-CSM2-MR	His/SSP245/SSP585	r1i1p1f1	320×160
3	CESM2	His/SSP245/SSP585	r1i1p1f1	288×192
4	CNRM-CM6-1	His/SSP245/SSP585	r1i1p1f2	256×128
5	CNRM-ESM2-1	His/SSP245/SSP585	r1i1p1f2	256×128
6	EC-Earth3-Veg	His/SSP245/SSP585	r1i1p1f1	512×256
7	FGOALS-g3	His/SSP245/SSP585	r1i1p1f1	180×80
8	GFDL-CM4	His/SSP245/SSP585	r1i1p1f1	288×180
9	INM-CM5-0	His/SSP245/SSP585	r1i1p1f1	180×120
10	MPI-ESM1-2-HR	His/SSP245/SSP585	r1i1p1f1	384×192
11	MRI-ESM2-0	His/SSP245/SSP585	r1i1p1f1	320×160

680

681 **Table 2.** Determination of “crossing years” for the periods reaching 1.5, 2 and 3 °C
 682 warming levels for different GCM and SSP combinations.

Models	1.5 °C warming level		2.0 °C warming level		3.0 °C warming level	
	SSP245	SSP585	SSP245	SSP585	SSP245	SSP585
ACCESS-ESM1-5	2024	2023	2037	2034	2070	2052
BCC-CSM2-MR	2026	2023	2043	2034	Not found	2054
CESM2	2024	2022	2037	2032	2069	2048
CNRM-CM6-1	2032	2028	2047	2039	2075	2055
CNRM-ESM2-1	2030	2026	2049	2039	2075	2058
EC-Earth3-Veg	2028	2023	2044	2035	2072	2053
FGOALS-g3	2033	2032	2063	2046	Not found	2069
GFDL-CM4	2025	2024	2038	2036	2073	2053
INM-CM5-0	2031	2027	2059	2038	Not found	2063
MPI-ESM1-2-HR	2032	2030	2055	2044	Not found	2066
MRI-ESM2-0	2024	2021	2038	2030	2074	2051

683

684 Table 3. Performance for CSSPv2 model simulations driven by the observed
 685 meteorological forcing (CMFD/CSSPv2) and the bias-corrected CMIP6 historical
 686 simulations (CMIP6_His/CSSPv2). The metrics include correlation coefficient (CC),
 687 root mean squared error (RMSE), and Kling-Gupta efficiency (KGE).

Variables			Experiments	CC	RMSE	KGE
Monthly	streamflow	over	CMFD/CSSPv2	0.95	165 m ³ /s	0.94
			CMIP6_His/CSSPv2	0.76	342 m ³ /s	0.71
Monthly	streamflow	over	CMFD/CSSPv2	0.93	169 m ³ /s	0.91
			CMIP6_His/CSSPv2	0.82	257 m ³ /s	0.81
Monthly	terrestrial	water	CMFD/CSSPv2	0.7	22 mm/month	-
	storage anomaly	over the	CMIP6_His/CSSPv2	0.4	24 mm/month	-
Sanjiangyuan region						
Annual	evapotranspiration		CMFD/CSSPv2	0.87	14 mm/year	-
		over the Sanjiangyuan region	CMIP6_His/CSSPv2	0.47	13 mm/year	-

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