

Intraseasonal variability of winter precipitation over Central Asia and the Western Tibetan plateau from 1979 to 2013 and its relationship with the North Atlantic Oscillation

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1 **Intraseasonal Variability of Winter Precipitation over Central Asia and the Western**
2 **Tibetan Plateau from 1979 to 2013 and its Relationship with the North Atlantic Oscillation**

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8 **Abstract**

9 Winter precipitation over Central Asia and the western Tibetan Plateau (CAWTP) is mainly
10 a result of the interaction between the westerly circulation and the high mountains around
11 the plateau. Empirical Orthogonal Functions (EOFs), Singular Value Decomposition (SVD),
12 linear regression and composite analysis were used to analyze winter daily precipitation and
13 other meteorological elements in this region from 1979 to 2013, in order to understand how
14 interactions between the regional circulation and topography affect the intraseasonal
15 variability in precipitation. The SVD analysis showed that the winter daily precipitation
16 variability distribution is characterized by a dipole pattern with opposite signs over the
17 northern Pamir Plateau and over the Karakoram Himalaya, similar to the second mode of EOF
18 analysis. This dipole pattern of precipitation anomaly is associated with local anomalies in
19 both the 700hPa moisture transport and the 500hPa geopotential height and is probably
20 caused by oscillations in the regional and large-scale circulations, which can influence the
21 westerly disturbance tracks and water vapor transport. The linear regression showed that the
22 anomalous mid-tropospheric circulation over CAWTP corresponds to an anti-phase variation

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23 of the 500 hPa geopotential height anomalies over the southern and northern North Atlantic
24 10 days earlier (at 95% significance level), that bear a similarity to the North Atlantic
25 Oscillation (NAO). The composite analysis revealed that the NAO impacts the downstream
26 regions including CAWTP by controlling south-north two branches of the middle latitude
27 westerly circulation around the Eurasian border. During the positive phases of the NAO, the
28 northern branch of the westerly circulation goes around the northwest Tibetan Plateau,
29 whereas the southern branch encounters the southwest Tibetan Plateau, which leads to a
30 reduced precipitation over the northern Pamir Plateau and an increased precipitation over
31 the Karakoram Himalaya, and vice versa.

32 KEY WORDS: Topographic precipitation; North Atlantic Oscillation; Westerly circulation;
33 Statistical analysis; Intraseasonal variability.

34

35 **1 Introduction**

36 The geographical region covered by Central Asia and the western Tibetan Plateau (CAWTP,
37 30–45° N, 60–85° E) has complex terrains and a unique climate. The Turan Depression is
38 located in the northwest, whereas the high mountains and plateaus (e.g. the Iran Plateau, the
39 Hindu Kush, the Karakoram Himalaya, the Pamir Plateau and the Tian Shan mountains) are
40 located from the southwest to the northeast (Fig. 1). The CAWTP region has an arid to semiarid
41 climate with an annual precipitation less than 400 mm, except for a few of the high mountain
42 areas. Because of the scarcity of water resources, there is a high risk that global climate change
43 will threaten both the natural environment and the human population in this region (Ragab &

44 Prudhomme 2002, WB et al. 2009).

45 Despite the complex topography and lack of meteorological stations, satellite remote
46 sensing data have been used to determine the spatiotemporal distribution of precipitation
47 over Central Asia (Guo et al. 2015) and the western Tibetan Plateau (Pohl et al. 2015). High-
48 resolution regional climate models have also been used to determine the patterns of
49 precipitation over CAWTP (Small et al. 1999, Schiemann et al. 2008, Ozturk et al. 2012,
50 Maussion et al. 2014).

51 Previous studies have reported the spatiotemporal distribution and regional differences
52 in precipitation over CAWTP and have found that the major weather system controlling the
53 winter precipitation over CAWTP is the westerly circulation (Schiemann et al. 2009, Yin et al.
54 2014). In winter, the westerly circulation transports moisture to Central Asia (Bothe et al.
55 2012), southwest Asia (Malik et al. 2015) and the western Tibetan Plateau (Curio et al. 2015).
56 The westerly circulation is disturbed by the high mountains in this region and causes heavy
57 precipitation and storms over the Pamir Plateau, the Hindu Kush, the Karakoram Himalaya and
58 the western Himalaya (Lang & Barros 2004, Cannon et al. 2015a, b). Yin et al. (2014) compared
59 the differences in precipitation climatology between the arid area of Central Asia and the East
60 Asia monsoon region and showed that winter is the rainy season in Central Asia, whereas the
61 rainy season occurs in summer for East Asia. They further showed that the control
62 atmospheric circulation over the western area changes between winter and summer. In winter
63 it is dominated by westerly upper-air flows, bringing moisture from the upstream to the region
64 while in summer it is dominated by northeasterly winds from the Asian interior, resulting in a

65 dry condition. The westerly circulation is not only the major weather system controlling the
66 winter mean precipitation over CAWTP, but also precipitation interannual variability and trend
67 in the region (Chen et al. 2011, Yin et al. 2014, Cannon et al. 2015a).

68 There have been many reports that large-scale atmospheric teleconnections regulate the
69 mid-latitude westerly circulation variability which in turn may influence interannual
70 precipitation variation over CAWTP (e.g. Aizen et al. 2001, Syed et al. 2006, Mariotti 2007,
71 Filippi et al. 2014, Yin et al. 2014, Cannon et al. 2015a, b, Hoell et al. 2015). Although some
72 evidence has been found that at interannual time scale the El Niño–Southern Oscillation may
73 be related to the precipitation variation in the cold season over Central and Western Asia
74 (Mariotti 2007, Hoell et al. 2015), other research has shown a close relationship between the
75 winter interannual precipitation variation over CAWTP and the North Atlantic Oscillation (NAO)
76 (Aizen et al. 2001, Syed et al. 2006, 2010, Yadav et al. 2009, Filippi et al. 2014, Yao and Chen
77 2015). Aizen et al. (2001) analyzed the relationship between mid-latitude precipitation in Asia
78 and the large-scale circulation of the atmosphere using data from hydro-meteorological
79 stations. Their results showed more precipitation over the Pamir and Tian Shan mountains
80 during the positive phases of the NAO. Syed et al. (2006) found a positive precipitation
81 anomaly over northwestern Asia that is well matched with the positive phase of the NAO.
82 Syed et al. (2010) presented a regional climate modeling study on both NAO and ENSO and
83 discussed the influence of westerly disturbances. Filippi et al. (2014) verified that winter
84 interannual precipitation variation over the Hindu Kush–Karakoram Himalaya region is
85 affected by the NAO. During the positive phases of the NAO, the Middle East, which is

86 upstream of CAWTP, experiences stronger westerly winds and evaporation, which leads to an
87 enhanced moisture transport and therefore enhanced precipitation over the Hindu Kush. Yao
88 and Chen (2015) found a significant negative correlation between the yearly precipitation on
89 mountains of the Syr Darya River Basin and NAO index from 1891 to 2011 but it turned out
90 non-significant on plains.

91 Previous studies have mainly considered the interannual variation in winter precipitation
92 over a specific area and its connection with the NAO and other large-scale circulations (e.g.
93 Aizen et al. 2001, Syed et al. 2006, Mariotti 2007, Filippi et al. 2014, Yin et al. 2014, Cannon et
94 al. 2015a, Hoell et al. 2015). However, little attention has been paid to the spatiotemporal
95 distribution of precipitation variability over CAWTP at the intraseasonal time scale. Therefore
96 it is important to investigate the spatial and temporal distribution of winter precipitation
97 variation over CAWTP on this time scale, to analyze its connection with regional and large-
98 scale circulations and to elucidate physical processes involved by using daily precipitation data
99 and other meteorological parameters.

100

101 **2 Data and methods**

102 Daily precipitation data from the Climate Prediction Center (Chen et al. 2008) and daily
103 meteorological variables of ERA-Interim from the European Centre for Medium-Range
104 Weather Forecasts (Dee et al. 2011) were used in this study. Sapna Rana (2017) have already
105 used the CPC Unified Rain gauge data and compared it with 9 other precipitation products
106 over the central southwest Asia. Their results show that the CPC data can reasonably reflect

107 the spatial-temporal distribution of winter precipitation over CAWTP despite some systematic
108 differences in the time means The used ERA-Interim variables include the daily 700 and 500
109 hPa geopotential height, meridional and zonal winds, the total column of water vapor (TCWV)
110 and the mean sea level pressure. The horizontal resolution of all the data is 0.5° longitude \times
111 0.5° latitude. In consideration of the higher accuracy of the reanalysis data obtained over high
112 mountain areas after 1979, when satellite data were first applied (Cannon et al. 2015b), we
113 chose all winters (December-March), consistent with other studies analyzing the relationships
114 between teleconnection patterns and precipitation in this area (Syed et al. 2006, 2010, Yadav
115 et al. 2009, Filippi et al. 2014) from 1979 to 2013, a total of 4244 days, as the study period.
116 The daily average data were obtained from four records with a six-hour interval for each day.
117 The encounter of westerly wind with high mountains of Tibetan Plateau happened around
118 700 hPa , which is closely connected with topographic precipitation, and therefore 700 hPa
119 wind is used to indicate low tropospheric circulation. The 500hPa geopotential height anomaly
120 can indicate the teleconnections.

121 To determine the characteristics of the winter intraseasonal precipitation and circulation
122 variability in CAWTP, the climatological seasonal cycle from December to the next March in
123 the daily data was removed before the analysis. We averaged the data of the same date of all
124 the years (1979-2013) and then calculated the 21 day moving mean as the climate mean state.
125 Thus, the daily values with climatological seasonal cycle removed were obtained by
126 subtracting the climate mean state from the actual data according to the following formula:

$$127 \quad Z_n^* = Z_n - \left(\sum_{i=-10,10} \left(\sum_{j=1,35} Z_{k+(i-1)\times 365+j} \right) / 35 \right) / 21$$

128 where n is the number of days from December 1, 1979; k is the number of days from
129 December 1 of a specific year; i is the number of years from 1979; j is the day from 10 days
130 before to 10 days after a specific day; and Z_n^* is the value after removing the climatological
131 seasonal cycle.

132 The EOF analysis was used to decompose winter daily precipitation variability over
133 CAWTP to find the spatial patterns as well as their time variation (Bjornsson and Venegas
134 1997). Figure 1 shows the study area with 1581 grid boxes and 4244 day time series at each
135 grid box, which is the basis for the EOF analysis.

136 Singular value decomposition (SVD) analysis (Wallace et al. 1992) was applied to show
137 the spatiotemporal relationship between the winter daily precipitation and the regional
138 circulation variations over CAWTP. This method identifies a pair of covaried spatial patterns,
139 their temporal variations and the covariance between two variables (Bjornsson and Venegas
140 1997).

141 Lead-lag linear regression was applied to study the connection between the 500 hPa
142 geopotential height variability over CAWTP and the upstream westerly jet variability (Wang
143 and Zhang 2015).

144 We also constructed composites of the precipitation distribution in CAWTP during
145 positive and negative NAO cases respectively, and discussed the relationship between the
146 NAO and the precipitation variability at the intraseasonal time scale. The NAO index was
147 defined as the standardized difference in mean sea level pressure between the southern
148 North Atlantic (25° – 40° N, 50° – 10° W) and the northern North Atlantic (50° – 65° N, 10° – 50° W)

149 (Hurrell 1996). After calculating the daily index for every winter from 1979 to 2013, the central
150 day of a positive (negative) NAO phase event was defined as the day with the relative
151 maximum (minimum) value of the NAO index in continuous five days and being greater (less)
152 than 1.5 (-1.5) at the same time. In this way, a total of 85 positive NAO phase cases and 82
153 negative phase cases were selected for the composites. Here we did composite analysis
154 based on cases with strong NAO anomalies in order to see teleconnections between two
155 regions and significant NAO related remote climatic anomalies over CAWTP. To trace westerly
156 wind disturbance (WWD) tracks, a wave-tracking approach was applied as documented in
157 Cannon (2015b) who defined the centers of the disturbances by standardized 500 hPa
158 geopotential height anomaly and a set of spatial threshold and temporal correlation to
159 identify the tracks. Here we used the 500hPa geopotential height anomaly of 1 standard
160 deviation and the spatial extent of 5 degrees as a set of thresholds to identify the location of
161 centers to count the WWDs. Then the WWD frequency in every grid was calculated.

162

163 **3 Spatial and temporal distribution of winter precipitation over CAWTP and its connection** 164 **with the contemporaneous regional circulation**

165 3.1 Winter precipitation climatology over CAWTP

166 The climate of Central Asia is controlled on an annual basis by westerly winds containing
167 little moisture, and it is classified as a typical arid to semiarid region. The rainy season usually
168 occurs in winter (from December to March) (Yin et al. 2014) and most of the moisture
169 transported by the strong westerly winds is intercepted to condense by the high mountains

170 of the region during this period (Syed et al. 2006). Figure 2a shows the spatial distribution of
171 the winter mean climatological precipitation over Asia from 1979 to 2013. There is more
172 precipitation over CAWTP and the main areas of precipitation are distributed on the windward
173 slope of the western Tibetan Plateau, in agreement with earlier reports by Schiemann et al.
174 (2008) and Yin et al. (2014). There are two core areas of precipitation centered at 70°E, 42°N
175 and 75°E, 33°N respectively, with the maximum precipitation rate of 8 mm d⁻¹. The winter
176 precipitation over CAWTP accounts for more than 50% of the annual precipitation (Fig. 2b).
177 The distribution of water vapor flux integrated vertically from 950 hPa to 300 hPa (Fig. 2c)
178 suggests that the moisture transport convergence over CAWTP is mainly controlled by mid-
179 tropospheric westerly winds, rather than by lower-tropospheric (below 950 hPa) circulations
180 (not shown). The distribution of WWD frequency (Fig. 2d) shows that more WWDs occur in
181 the Karakoram Himalaya, coincident with the local precipitation maximum there shown in Fig.
182 2a. This is consistent with Cannon et al. (2015a, b) who reported that heavy precipitation
183 occurs when the WWD encountered mountains of the Himalaya. The climatology of
184 precipitation over CAWTP is quite different from that over the monsoon areas. The annual
185 precipitation over CAWTP is much lower than that over Asian monsoon regions and mainly
186 occurs in winter, when the mountains block and lift the strong mid-latitude westerly
187 circulation (e.g., Yin et al 2014, Cannon et al. 2015a, b).

188 189 3.2 Intraseasonal variability in winter daily precipitation and circulation over CAWTP

190 It is possible that the strength and position of westerly circulation determine the forced

191 upward motion, frequency of WWDs and the distribution of precipitation on the intraseasonal
192 time scale. The winter daily precipitation was decomposed by EOF analysis after the
193 climatological seasonal cycle was removed and the first two modes are shown in Fig. 3a and
194 3b. The first mode shows the same sign all over CAWTP with explained variance of 24.4% and
195 the second mode shows a seesaw pattern between northern Pamir Plateau and the
196 Karakoram Himalaya with explained variance of 15.6%. These EOF modes suggest the
197 existence of the intraseasonal variability of winter precipitation over CAWTP. Then we
198 composited the 500 hPa zonal wind (U component) during the dominant periods according to
199 the time series of the first mode and the second mode as Fig. 3c and Fig. 3d, which indicated
200 close connection between regional circulation and precipitation. Therefore, the relationship
201 between the regional circulation and precipitation in the study area is explored in the
202 following.

203 The seasonal cycle removed daily precipitation and 700hPa moisture flux fields for 35
204 winters during 1979–2013 were used for the SVD analysis to determine the simultaneous
205 relationship between the precipitation and circulation variability over CAWTP. The first SVD
206 mode shows a dipole pattern of precipitation variability with a positive anomaly over the
207 Karakoram Himalaya (center located at 33°N, 78°E, Fig.4a) and a negative anomaly over the
208 northern Pamir Plateau (center located at 40°N, 68°E), bearing a very similar structure as
209 shown for the second mode of EOF analysis (Fig. 3b). Accompanied with this pattern of
210 precipitation variability is a southerly wind anomaly over the Karakoram Himalaya and a
211 northwesterly wind anomaly (southwesterly wind weakened) over the northern Pamir

212 Plateau(not shown) and anomalous lower tropospheric moisture transport (Fig. 4b). This
213 pattern of precipitation anomaly is closely related to the forced lifting of the westerly
214 airstream on the windward side of the mountains. The explained variance reaches 55.12% and
215 the correlation coefficient between the time coefficients of two fields is 0.65 for the first SVD
216 mode, suggesting the importance of orographic forcing in modulating the regional
217 precipitation at the intraseasonal time scale.

218 We also did the SVD analysis with the seasonal cycle removed daily precipitation and 500
219 hPa geopotential height fields over CAWTP. The explained variance and correlation coefficient
220 corresponding for the first SVD mode are 60.5% and 0.64, respectively (Fig. 4c and d). Figs. 4c
221 and 4d illustrate that a negative center over the southeast of CAWTP in the field of 500 hPa
222 geopotential height anomaly is associated with a precipitation anomaly pattern with the
223 wetter southeast and the drier northwest. Hence, the distribution of winter precipitation
224 anomaly in the study area at the intraseasonal time scale is the result of topography-affected
225 regional atmospheric circulation variability. In the following, a lead-lag linear regression was
226 used to further reveal the relationship between the regional and large-scale circulation
227 variations and the regional precipitation change.

228

229 **4 Correlation of 500hPa geopotential height over CAWTP with the preceding atmospheric** 230 **circulation over the North Atlantic Ocean and Eurasia**

231 The preceding atmospheric circulation over the Atlantic Ocean and Eurasian region may
232 influence the regional circulation over CAWTP, leading to the regional precipitation anomaly

233 via teleconnection. The time series of the 500 hPa geopotential height analyzed by SVD was
234 normalized and then the normalized time series with an absolute value greater than 1.5 and
235 higher than the neighboring 4 other days was chosen for further analysis (a total of 200 time
236 series mixed with both positive and negative values). A linear regression was applied between
237 the chosen 200 time series and the antecedent 500 hPa geopotential height field over the
238 Atlantic Ocean–Eurasian region from lead 16 days to the lag 3 day of the chosen days
239 respectively and results are shown in figure 5. There is a positive geopotential height anomaly
240 over the Azores (center located at 30°W and 40°N) and a negative anomaly over Iceland
241 (center located at 10°W and 60°N) from day -10 to day -8 (10 to 8 days before the regional
242 circulation phase, the same below) (Fig. 5a, 5b). These patterns of the anomalous circulation
243 show a similarity to the positive phase of the NAO (Hurrell 1996). On day -6, the original
244 negative geopotential height anomaly over Iceland moves to the east and a significant positive
245 anomaly occurs over the Eurasian border and northern Asia and meanwhile a negative
246 anomaly over CAWTP develops (Fig. 5c). With time advance, the magnitude of the anomaly
247 over the Atlantic Ocean decrease, whereas the strength of the anomalies over the Eurasian
248 border and over CAWTP enhance on day -4 (Fig. 5d). On day -2 and 0, significant negative
249 geopotential height anomalies occur over CAWTP (Fig. 5e, 5f), similar to the results of the SVD
250 analysis. It is therefore possible that the regional circulation has been affected by the
251 preceding large-scale circulation variability related to NAO.

252 The time evolution of the 500hPa geopotential height anomaly over the North Atlantic
253 was analyzed by lead-lag linear regression and correlation. Areas A (20–40° W, 30–40° N), B

254 (0–20° W, 54–64° N), C (44–64° E, 54–64° N) and D (60–80° E, 30–40° N) in Fig. 5 were chosen
255 to calculate the average geopotential height anomaly and the correlation coefficient in
256 different lead days. The four domains are recognized in consideration of both covering the
257 centers of 500hPa geopotential height anomaly and the significant at 95% level in Fig. 5.
258 Otherwise, the numbers of the grids of the four domains should be the same so they are
259 comparable. Table 1 lists regression coefficients and correlation coefficients of regionally-
260 averaged 500 hPa geopotential heights on different lag days for these four areas and the time
261 coefficients of the first mode of 500 hPa geopotential height field over CAWTP in the above
262 SVD analysis. From days –15 to –9 (the negative number means the days before the day for
263 the regional SVD analysis) there is a positive anomaly over area A, a significant increasingly
264 negative anomaly over area B, and no significant anomaly over area C. From days –8 to –4,
265 the anomaly over area A decreases and the anomaly over area B increases to a maximum
266 value. A significant positive anomaly appears over area C on day –5 and increases rapidly
267 afterwards. From days –3 to 0 there is no significant anomaly over area A, the anomaly over
268 area B decreases, and the anomaly over areas C and D increases to maximum values. From
269 days +1 to +3 there is no significant anomaly over area B and the anomaly over areas C and D
270 decreases. Figure 5 and Table 1 show time evolutions of geopotential height anomaly in
271 various regions before the maximum height anomaly over CAWTP and suggest that NAO-like
272 circulation anomaly over the North Atlantic is a precursor for the intraseasonal variation of
273 precipitation over CAWTP. The lead-lag correlation coefficients between NAO index and time
274 series of first SVD mode in these chosen days were calculated and they are given in Table 1.

275 These correlation coefficients indicate large correlations when NAO leads the regional
276 geopotential height by about 10 days.

277 Although we found a close connection between the 500 hPa geopotential height over
278 CAWTP and geopotential height over the North Atlantic about 10-8 days before by linear
279 regression and correlation, further analysis was required to elucidate the mechanism for the
280 relationship between the north–south precipitation seesaw over CAWTP during both positive
281 and negative NAO phases respectively. In the following, composite analysis was therefore
282 used to further investigate how the different NAO phases influenced the regional circulation
283 and precipitation over CAWTP.

284

285 **5 Influence of the NAO on spatial and temporal distribution of winter precipitation over** 286 **CAWTP**

287 Based on our definition of NAO events, the 500 hPa geopotential height, the 700 hPa wind
288 flow field were combined to assess the influence of the NAO on the circulation over remote
289 downstream areas. The time evolutions of composited 500hPa geopotential height anomaly
290 and 700 hPa streamlines at different lags for the positive and negative NAO phase cases are
291 illustrated in Fig. 6. In positive NAO phases, the NAO pattern is clearly shown on the 0 day (the
292 peak day of NAO index) with the strong negative 500 hPa geopotential height anomaly over
293 the Iceland and the positive one over the Azores. As time advances, the positive anomaly
294 extends eastward to the Eurasian border. In the following days, NAO pattern is weakening and
295 the east positive anomaly is further extended eastward and is finally separated from the

296 positive anomaly over the Azores. The eastward extension of positive anomaly over the
297 Eurasian border is associated with the intensification of the negative anomaly over the CAWTP
298 (Fig. 6c). The time evolutions of the 700 hPa wind streamlines associated with geopotential
299 height anomaly evolutions offer another view. Associated with positive NAO at day 0 is a single
300 strong westerly jet over the North Atlantic and two branches of strong westerly wind, whose
301 axes locate on 55°N and 30°N along the Eurasian border, resulted from the positive 500 hPa
302 geopotential height anomaly over the Eurasian border (Fig. 6a). Over the Eurasian border, the
303 north branch westerly wind forms a ridge while the south one forms a trough. They both
304 extend eastward in a similar way as the positive 500 hPa geopotential height anomaly. The
305 north branch westerly wind goes around the north CAWTP while the south branch encounters
306 the Karakoram Himalaya. The Figs. 6d, 6e and 6f show time evolutions of composited 500hPa
307 geopotential height anomaly and 700 hPa streamlines at different days for the negative NAO
308 phases. Time evolutions of geopotential height anomaly show similar evolutions as those
309 during the positive NAO phase cases, but with a change in sign of the anomalies. Associated
310 with the negative NAO phases, the north branch of westerly wind encounters the north
311 CAWTP while the south branch goes around the Krakoram-Himalaya at day 10.

312 The time averaged 700 hPa moisture flux anomaly and WWD frequency anomaly from
313 day 5 to day 10 after the peak of NAO phases intuitively show the regulation of regional
314 circulation by the NAO. During the positive NAO phases, northeasterly moisture flux
315 anomalies (decreased southwesterly moisture flux) occur over the Central Asia and northern
316 Pamir Plateau, whereas southerly moisture flux anomalies accompanied with higher

317 frequency of WWD occur over the Karakoram Himalaya (Fig. 7a). During the negative NAO
318 phases, southwesterly moisture flux anomalies occur over the Central Asia and northern
319 Pamir Plateau, whereas northeasterly moisture flux anomalies accompanied with lower
320 frequency of WWD occur over the Karakoram Himalaya (Fig. 7b). The composites of
321 precipitation anomalies of positive and negative NAO phases, as shown in Fig. 7c and 7d, are
322 well matched with the regional circulation and moisture flux anomalies. There is less
323 precipitation over the north east of Iran, Turkmanistan, Tajikistan and Kyrgyzstan (center at ~
324 75°E, 34°N) and more precipitation over the Karakoram Himalaya (center at ~ 70°E, 42°N)
325 during positive NAO phases and vice versa. For the south part, the WWD seems to be more
326 important while the moisture flux takes control of the north part. Compared with the daily
327 average winter precipitation (Fig. 2a), the precipitation anomaly at the intraseasonal time
328 scale reaches 10% of the climatological mean. The precipitation composites during the
329 different NAO phases are similar to the results of the SVD analyses. The two analyses both
330 demonstrate the teleconnection between the NAO over the North Atlantic and the regional
331 circulation and precipitation variability over CAWTP.

332

333 **6 Discussion and conclusions**

334 The climate is characterized by dominant westerly circulation and scarce precipitation
335 over CAWTP. The lowlands of this area are almost rainless due to the lack of moisture and
336 upward motion. By contrast, heavy rains occur on the high mountain areas as a result of the
337 uplifting of the westerly winds by the high topography.

338 There have been much published research on precipitation and its interannual variability
339 over the Karakoram Himalaya (Syed et al. 2006, 2010, Filippi et al. 2014, Cannon et al. 2015a,
340 b). For example, Cannon et al. (2015a, b) stressed the importance of the westerly disturbance.
341 Less research has focused on precipitation over the northern Pamir Plateau and mainly
342 focuses on interannual variations (e.g., Chen et al. 2011). Filippi et al. (2014) and Syed et al.
343 (2006) found that the interannual variation in precipitation is connected to the NAO, but they
344 did not discuss the intraseasonal precipitation variations and the contemporaneous opposite
345 variations over the northern Pamir Plateau and over the Karakoram Himalaya.

346 This study focused on the intraseasonal variation in winter daily precipitation over CAWTP.
347 The EOF, SVD, lead-lag linear regression and composite analyses showed that a seesaw pattern
348 of winter intraseasonal precipitation anomaly between the Karakoram Himalaya and the
349 northern Pamir Plateau is connected with the intraseasonal oscillation of the NAO. The main
350 physical processes for the seasaw pattern of precipitation variability are the trough–ridge
351 phase of the two westerly jets rather than the westerly strength at the intraseasonal scale.
352 Multi-method analyses gave similar results, confirming that the NAO is able to influence the
353 intraseasonal precipitation variability in winter over CAWTP.

354 The seesaw pattern of winter intraseasonal precipitation anomaly was found between
355 the southeastern CAWTP (centered in the Karakoram Himalaya) and the northwestern CAWTP
356 (centered in the northern Pamir Plateau) by the SVD analysis and this pattern of precipitation
357 was closely connected to the regional 700 hPa circulation. When there was a northeasterly
358 moisture flux anomaly (southwesterly moisture flux weakened) over the northwestern CAWTP

359 and a southwesterly moisture flux anomaly (southwesterly moisture flux strengthened) over
360 the southwestern CAWTP at 700 hPa, a negative precipitation anomaly occurred over the
361 northern Pamir Plateau and a positive anomaly over the Karakoram Himalaya. Accompanied
362 by the seesaw precipitation pattern, a negative 500 hPa geopotential height anomaly over
363 CAWTP (center located in the central Tibetan Plateau) was also observed, and vice versa.

364 The mechanism for the seesaw precipitation anomaly over CAWTP may be related to the
365 NAO-like circulation at the intraseasonal time scale. The results of both the regression and
366 composite analyses showed that the seesaw precipitation pattern was closely connected with
367 the precursor NAO-like circulation anomalies over the North Atlantic. During the positive NAO
368 phases, the southern branch of the 700 hPa westerly winds formed a ridge over the North
369 Atlantic and combined with the northern branch which formed a trough. The two branches
370 separated around Eurasian border. The southern branch then formed a trough over the Middle
371 East, which increased the transport of moisture and the strength of the westerly winds over
372 the Karakoram Himalaya. By contrast, the northern branch formed a ridge and went around
373 the northwestern Tibetan Plateau, which decreased southwesterly winds, the southwesterly
374 moisture transport, and therefore precipitation over the Central Asia and northern Pamir
375 Plateau. The large scale wind stream pattern can modify the regional circulation. Meanwhile
376 during positive NAO, more WWD occurred over the Karakoram Himalaya since south branch
377 of westerly encounters the Himalaya, which finally increased the precipitation over Karakoram
378 Himalaya. This pattern was reversed during the negative NAO phases. The main physical
379 processes involved and discussed above for the influence of the NAO on the winter

380 precipitation variability at the intraseasonal time scale over CAWTP are illustrated in
381 schematic diagrams of Fig. 8.

382

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LITERATURE CITED

392 Aizen E M, Aizen V B, Melack J M and others (2001) Precipitation and atmospheric circulation
393 patterns at mid-latitudes of Asia. *International Journal of Climatology* 21: 535-556

394 Bjornsson H, Venegas S A (1997) A manual for EOF and SVD analyses of climatic data.
395 CCGCR Report97: 112-134

396 Bothe O, Fraedrich K, Zhu X (2012) Precipitation climate of Central Asia and the large-scale
397 atmospheric circulation. *Theoretical and Applied Climatology*108: 345-354

398 Cannon F, Carvalho L M V, Jones C and others (2015a) Multi-annual variations in winter
399 westerly disturbance activity affecting the Himalaya. *Climate Dynamics* 44: 441-455

400 Cannon F, Carvalho L M V, Jones C and others (2015b) Winter westerly disturbance dynamics

401 and precipitation in the western Himalaya and Karakoram: a wave-tracking approach.
402 Theoretical and Applied Climatology: 1-18

403 Chen FH, Huang W, Jin LY, Chen J and others (2011) Spatiotemporal precipitation variations
404 in the arid Central Asia in the context of global warming. *Science China Earth Sciences*
405 54: 1812–1821

406 Chen M, Shi W, Xie P and others (2008) Assessing objective techniques for gauge-based
407 analyses of global daily precipitation. *Journal of Geophysical Research: Atmospheres* 113:
408 D04110, doi:10.1029/2007JD009132

409 Curio J, Maussion F, Scherer D (2015) A 12-year high-resolution climatology of atmospheric
410 water transport over the Tibetan Plateau. *Earth System Dynamics* 6: 109-124

411 Dee D P, Uppala S M, Simmons A J and others (2011) The ERA-Interim reanalysis:
412 Configuration and performance of the data assimilation system. *Quarterly Journal of the*
413 *Royal Meteorological Society* 137: 553-597

414 Filippi L, Palazzi E, von Hardenberg J and others (2014) Multidecadal variations in the
415 relationship between the NAO and winter precipitation in the Hindu Kush–Karakoram.
416 *Journal of Climate* 27: 7890-7902

417 Guo H, Chen S, Bao A and others (2015) Inter-comparison of high-resolution satellite
418 precipitation products over Central Asia. *Remote Sensing* 7: 7181-7211

419 Hoell A, Shukla S, Barlow M and others (2015) The forcing of monthly precipitation variability
420 over Southwest Asia during the boreal cold season. *Journal of Climate* 28: 7038-7056

421 Hurrell, J W (1996) Decadal trends in the North Atlantic Oscillation: regional temperatures

422 and precipitation. *Geophysical Research Letters* 23: pp. 665–668

423 Lang T J, Barros A P (2004) Winter storms in the central Himalayas. *Journal of the*
424 *Meteorological Society of Japan*82: 829-844

425 Malik K M, Taylor P A, Szeto K (2015) Characteristics of moisture flux convergence in Central
426 Southwest Asia. *Theoretical and Applied Climatology* 120: 643-659

427 Mariotti A (2007) How ENSO impacts precipitation in southwest central Asia. *Geophysical*
428 *Research Letters*34:L16706, doi:10.1029/2007GL030078

429 Maussion F, Scherer D, Mölg T and others (2014) Precipitation seasonality and variability over
430 the Tibetan Plateau as resolved by the High Asia reanalysis. *Journal of Climate*27: 1910-
431 1927

432 Ozturk T, Altinsoy H, Türkes M and others (2012) Simulation of temperature and precipitation
433 climatology for the Central Asia CORDEX domain using RegCM 4.0. *Climate*
434 *Research*52: 63-76

435 Pohl E, Gloaguen R, Seiler R (2015) Remote sensing-based assessment of the variability of
436 winter and summer precipitation in the Pamirs and their effects on hydrology and hazards
437 using harmonic time series analysis. *Remote Sensing*7: 9727-9752

438 Ragab R, Prudhomme C (2002) Climate change and water resources management in arid and
439 semi-arid regions: prospective and challenges for the 21st century. *Biosystems*
440 *Engineering* 81: 3-34

441 Rana S, McGregor J, Renwick J (2017) Wintertime precipitation climatology and ENSO
442 sensitivity over central southwest Asia. *International Journal of Climatology* 37: 1494-

443 1509

444 Schiemann R, Lüthi D, Vidale P L and others (2008) The precipitation climate of Central
445 Asia—intercomparison of observational and numerical data sources in a remote semiarid
446 region. *International Journal of Climatology*28: 295-314

447 Schiemann R, Lüthi D, Schär C (2009) Seasonality and interannual variability of the westerly
448 jet in the Tibetan Plateau region. *Journal of Climate* 22: 2940-2957

449 Small E E, Giorgi F, Sloan L C (1999) Regional climate model simulation of precipitation in
450 central Asia: Mean and interannual variability. *Journal of Geophysical Research:*
451 *Atmospheres*104(D6): 6563-6582

452 Syed F S, Giorgi F, Pal J S and others (2006) Effect of remote forcings on the winter
453 precipitation of central southwest Asia part 1: observations. *Theoretical and Applied*
454 *Climatology*86: 147-160

455 Syed F S, Giorgi F, Pal J S, et al. (2010) Regional climate model simulation of winter climate
456 over Central–Southwest Asia, with emphasis on NAO and ENSO effects. *International*
457 *Journal of Climatology*30(2): 220-235

458 Wallace J M, Smith C, Bretherton C S (1992) Singular value decomposition of wintertime sea
459 surface temperature and 500-mb height anomalies. *Journal of Climate*5: 561-576

460 Wang N, Zhang Y (2015) Connections between the Eurasian teleconnection and concurrent
461 variation of upper-level jets over East Asia. *Advances in Atmospheric Sciences*32: 336-
462 348

463 WB, ISDR, CAREC (2009) Central Asia and Caucasus disaster risk management initiative

464 (CAC DRMI), Risk assessment for Central Asia and Caucasus. Desk study review: 1-
465 155.[http://www.preventionweb.net/files/11641_CentralAsiaCaucasusDRManagementIni](http://www.preventionweb.net/files/11641_CentralAsiaCaucasusDRManagementInit.pdf)
466 t.pdf (accessed 15 Sep 2016)

467 Yadav R K, Rupa Kumar K, Rajeevan M. (2009) Increasing influence of ENSO and decreasing
468 influence of AO/NAO in the recent decades over northwest India winter precipitation.
469 Journal of Geophysical Research: Atmospheres 114(D12)

470 Yao J, Chen Y. (2015) Trend analysis of temperature and precipitation in the Syr Darya Basin
471 in Central Asia. Theoretical and Applied Climatology 120(3-4): 521-531

472 Yin ZY, Wang HL, Liu XD (2014) A comparative study on precipitation climatology and
473 interannual variability in subtropical East Asia and Central Asia. Journal of Climate 27:
474 7830-7848
475

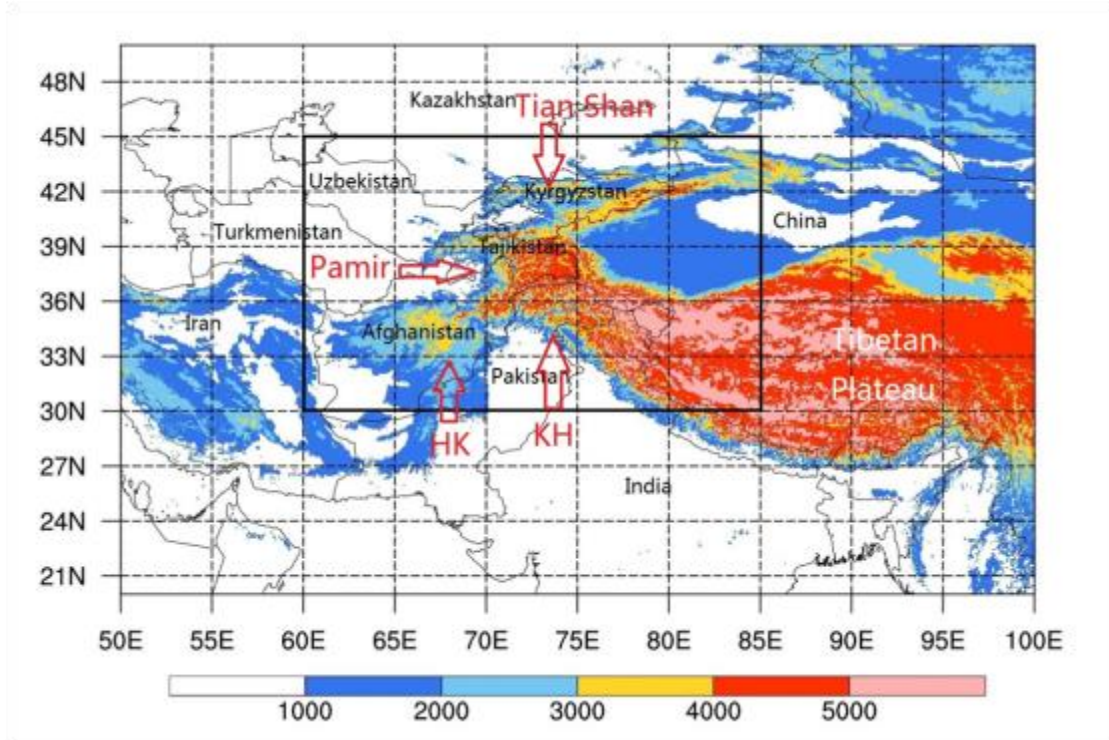
476 Table 1 Correlation coefficients of regionally-averaged 500 hPa geopotential heights on
477 different lead-lag days for four regions indicated in Fig. 5 with the time coefficients of the
478 first mode of 500 hPa geopotential height field over CAWTP in the SVD analysis mentioned in
479 the text, and the correlation coefficients between NAO index and the time series. A negative
480 lag means that the regionally-averaged height value over the four regions leads the height
481 anomaly over CAWTP in the SVD analysis.

Lag day	Area A	Area B	Area C	Area D	NAOI
-16	0.13	-0.14	-0.06	-0.15*	0.19**
-15	0.16*	-0.22**	-0.01	-0.20**	0.22**
-14	0.17*	-0.23**	0.02	-0.24**	0.24**
-13	0.15*	-0.24**	0.04	-0.26**	0.24**
-12	0.14	-0.30**	0.07	-0.29**	0.23**
-11	0.13	-0.33**	0.11	-0.30**	0.21**
-10	0.15*	-0.30**	0.11	-0.32**	0.20**
-9	0.17*	-0.32**	0.11	-0.35**	0.20**
-8	0.14	-0.36**	0.10	-0.38**	0.21**
-7	0.13	-0.44**	0.11	-0.41**	0.20**
-6	0.17*	-0.48**	0.16	-0.44**	0.16*
-5	0.20**	-0.46**	0.26**	-0.47**	0.12
-4	0.17*	-0.46**	0.39**	-0.51**	0.08
-3	0.14	-0.40**	0.54**	-0.59**	0.05
-2	0.14	-0.34**	0.67**	-0.68**	0.04
-1	0.16*	-0.23**	0.70**	-0.75**	0.04
0	0.15*	-0.17*	0.67**	-0.67**	0.00
1	0.12	-0.16*	0.59**	-0.57**	-0.08
2	0.08	-0.06	0.48**	-0.47**	-0.14
3	0.05	-0.03	0.44**	-0.39**	-0.14

482 *Significant at the 95% level

**Significant at the 99% level

483

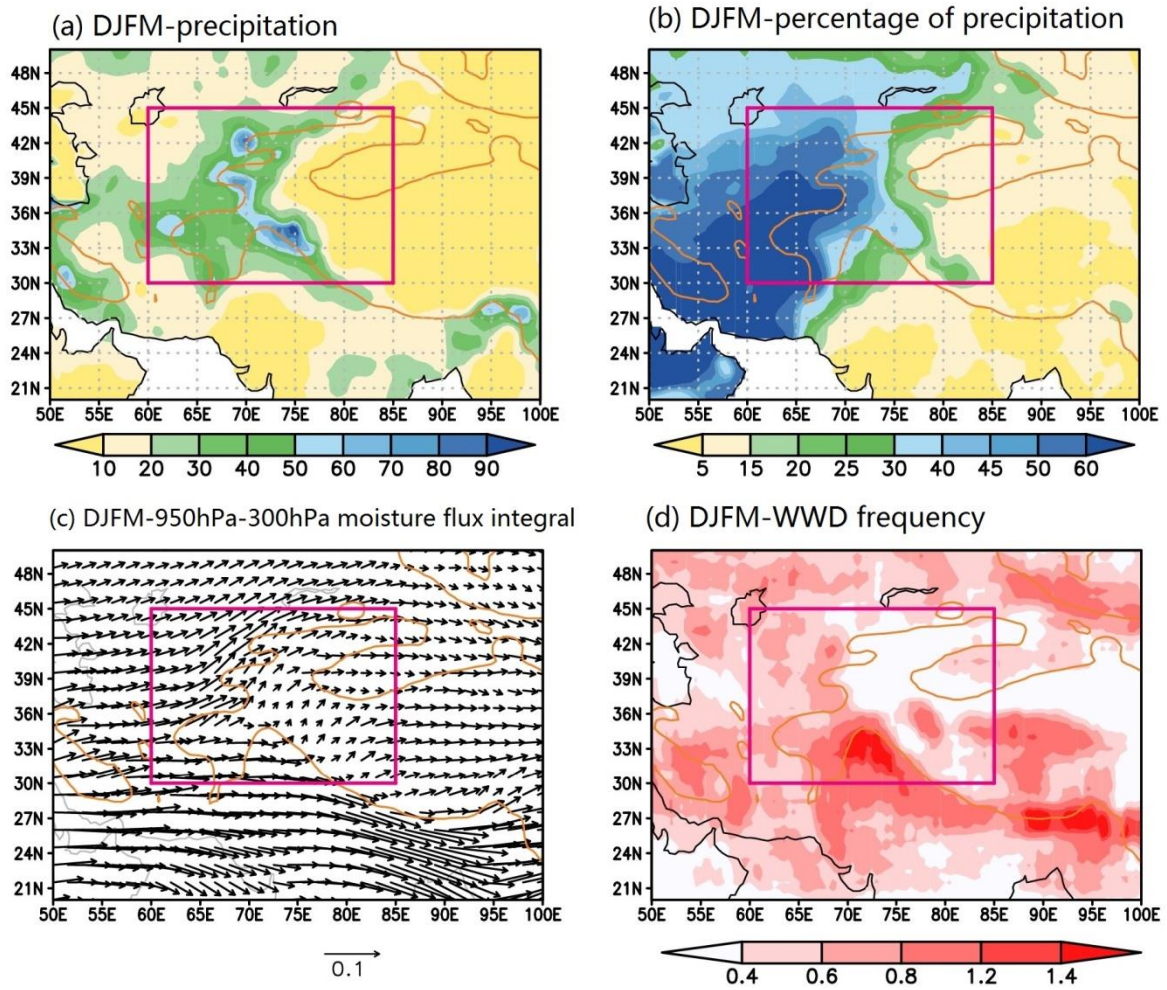


484

485 Fig. 1. Topography (color shaded areas, in meters) and national boundaries (thin gray lines) of

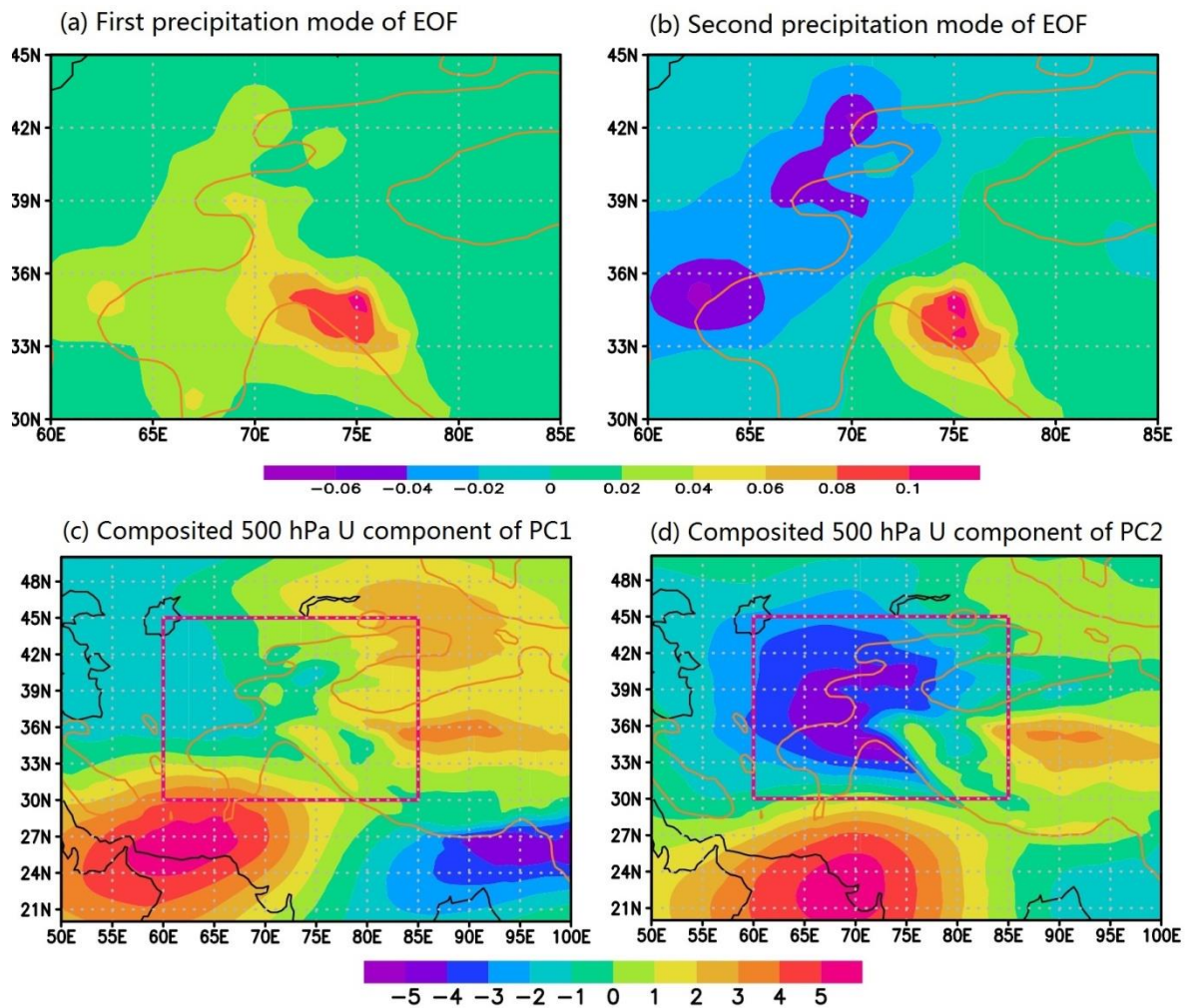
486 Central Asia, the western Tibetan Plateau (outlined by the black box) and surrounding areas.

487 HK and KH indicate Hindu Kush and Karakoram Himalaya, respectively.



488

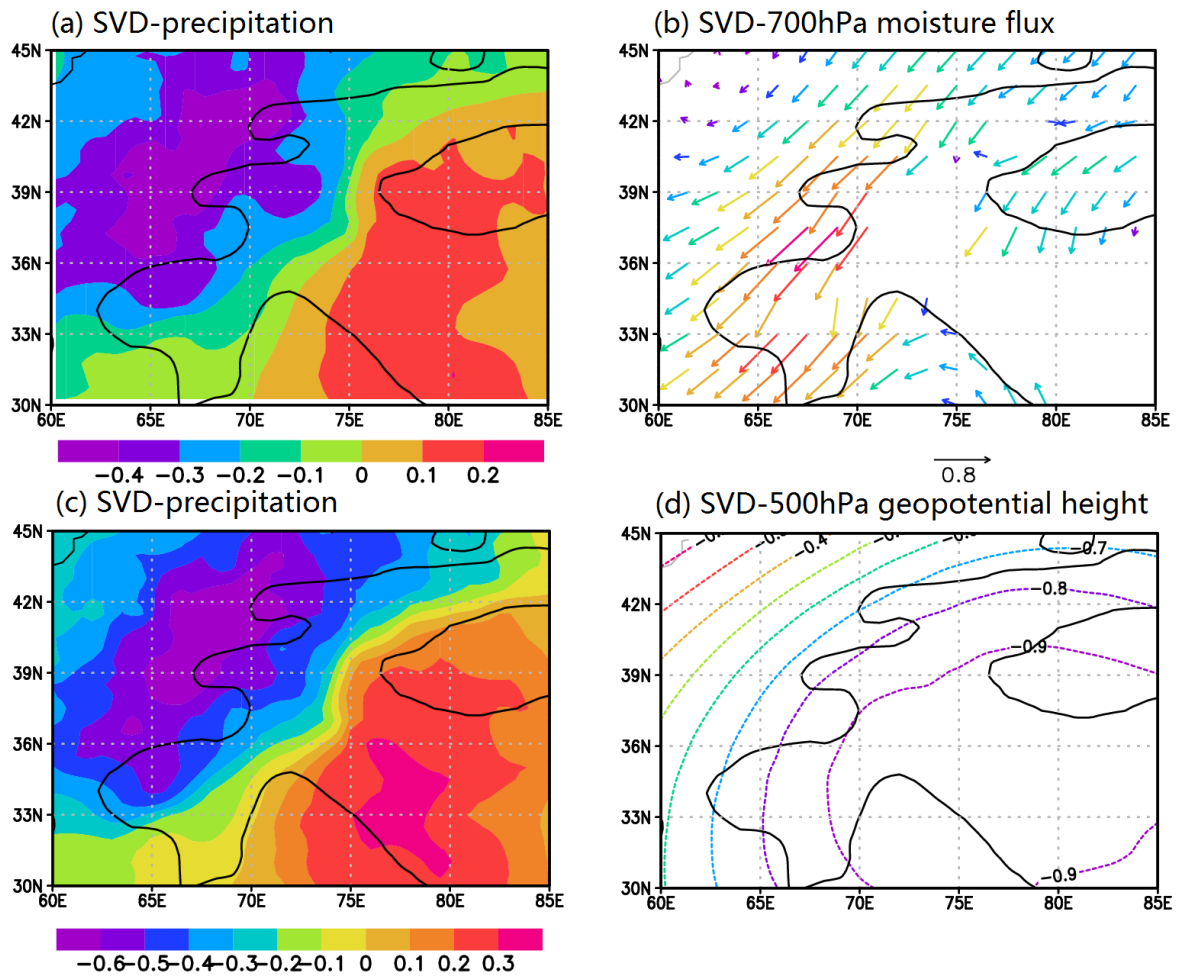
489 Fig. 2. (a) Average precipitation (0.1 mm d^{-1}), (b) percentage of annual precipitation falling in
 490 winter (%), (c) water vapor flux integrated vertically for 950–300 hPa ($\text{kg m}^{-1} \text{ s}^{-1}$) and (d)
 491 westerly wind disturbances frequency (percentage per day) in the CAWTP and surrounding
 492 areas in winter (December–March) for 1979–2013. The brown lines indicate the topographic
 493 contour of 1500 m and the red boxes represent the CAWTP region.



494

495 Fig. 3. First mode (a) and second mode (b) of the EOF analysis of daily precipitation field for
 496 winters during 1979-2013. The composited 500 hPa zonal wind (U component) of the
 497 dominant periods of the First mode (c) and second mode (d). The brown lines represent the
 498 topographic contour of 1500 m.

499



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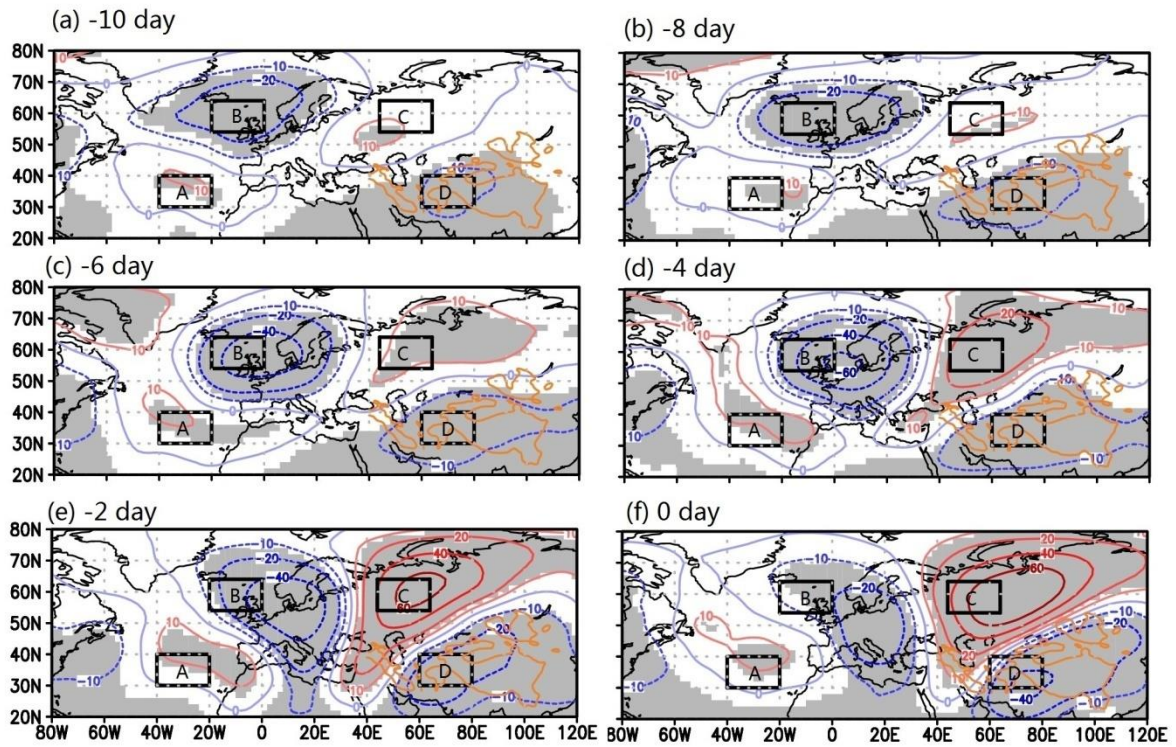
502 Fig. 4. First mode of the SVD analysis between seasonal cycle removed daily precipitation field

503 (a) and 700 hPa daily moisture flux field (b) for winters during 1979-2013. Panels (c) and (d)

504 are the same as (a) and (b), respectively, but for precipitation field (c) and 500 hPa

505 geopotential height field (d). The black lines represent the topographic contour of 1500 m.

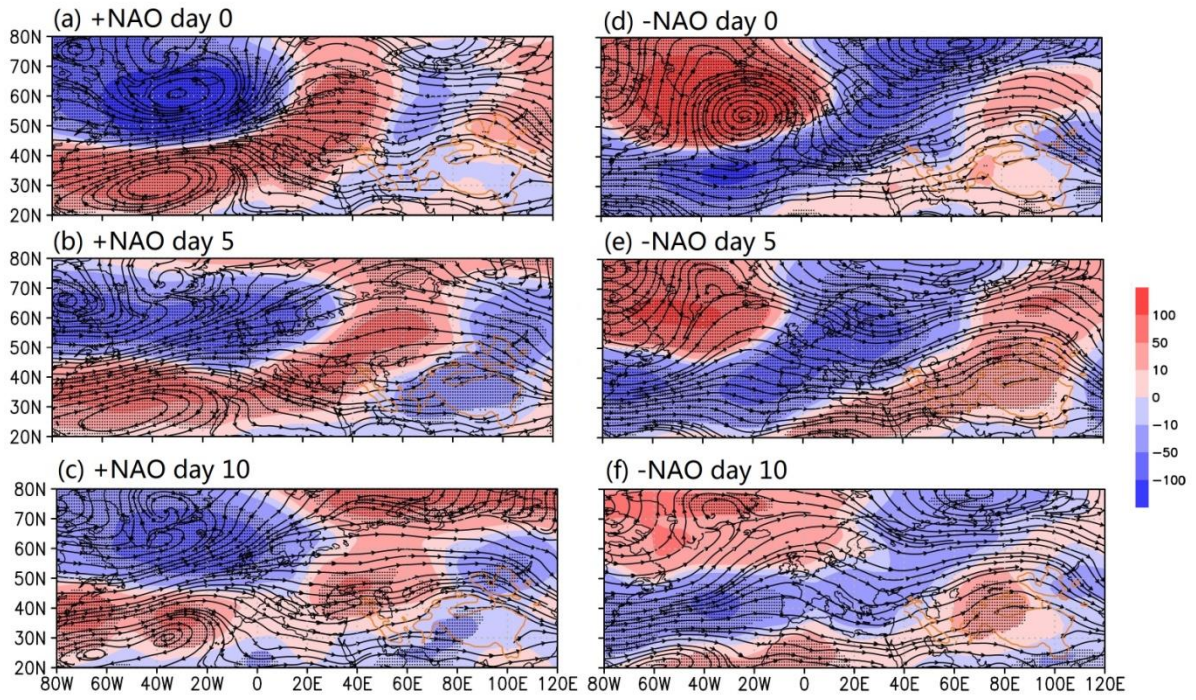
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507

508 Fig. 5. Regressed 500hPa geopotential height anomaly fields for days (a) -10, (b) -8, (c) -6, (d)
 509 -4, (e) -2 and (f) 0 with the chosen 200time coefficients of the first mode of 500 hPa
 510 geopotential height field over CAWTP in the SVD analysis. The negative value indicates the
 511 number of days before the day of SVD analysis. The contours represent the regression
 512 coefficients and the gray shading represents the areas that were significant at the 95% level.
 513 The brown lines represent the 1500 m topographic height and the black boxes represent the
 514 four geopotential height anomaly centers.

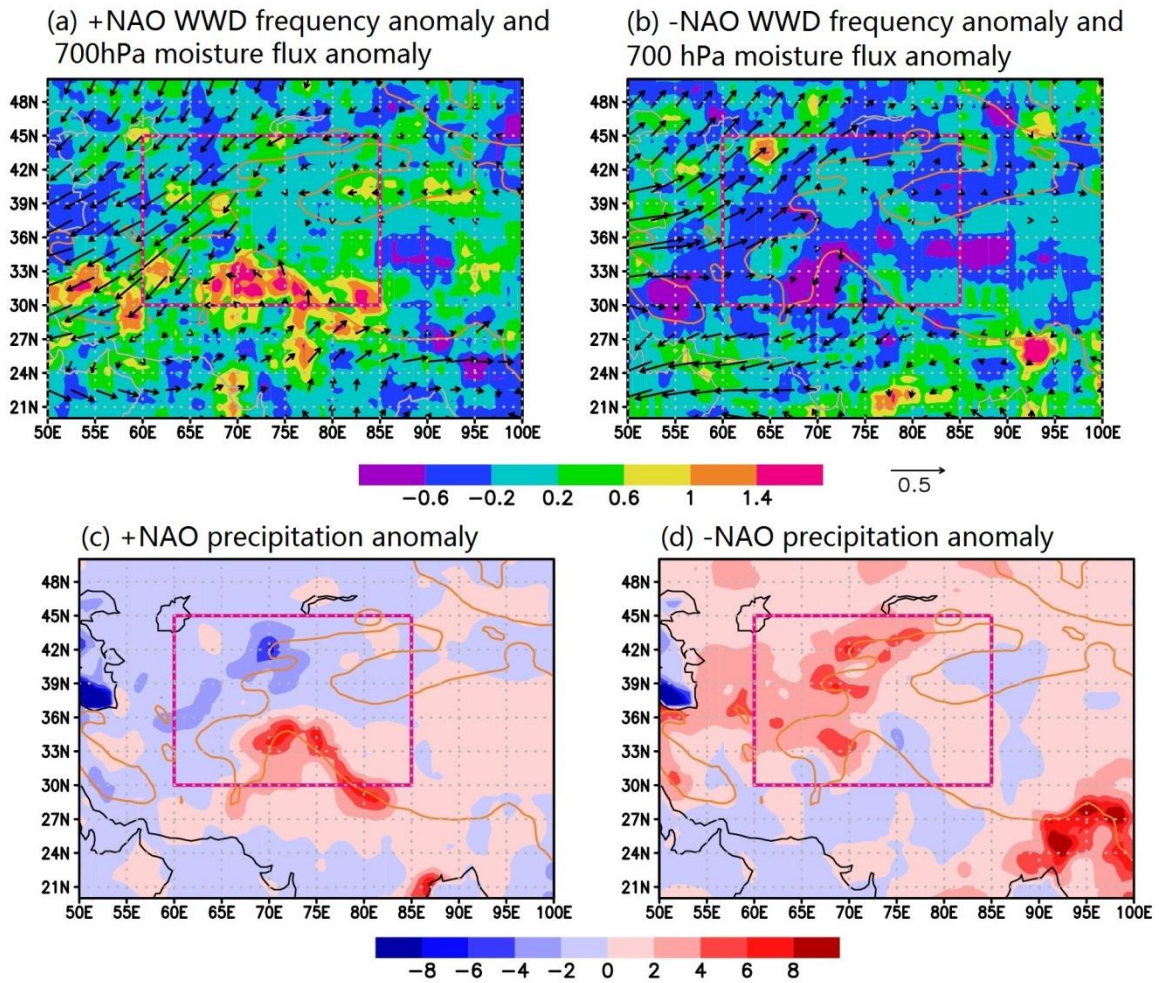
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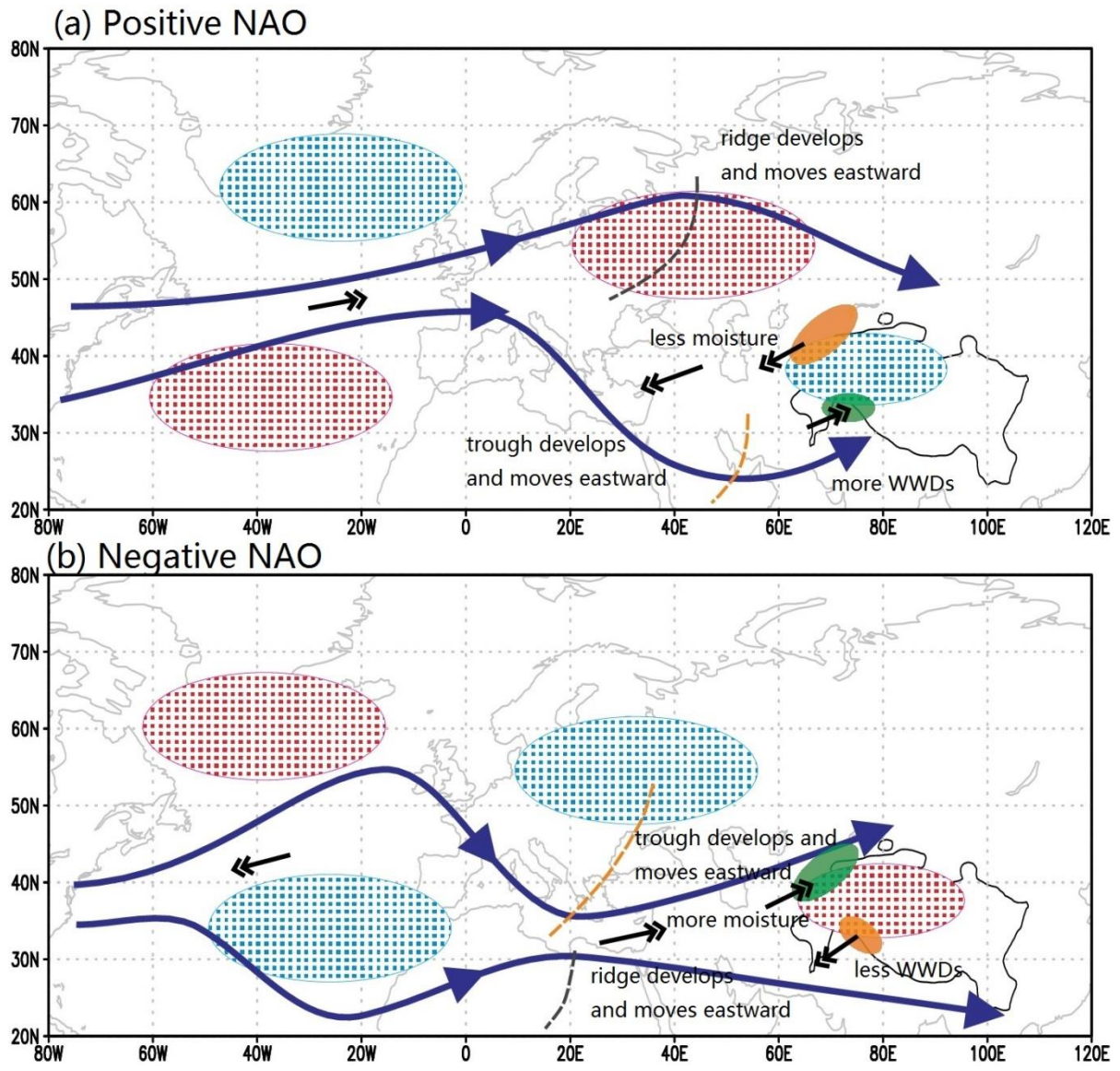
517 Fig. 6. The 500 hPa geopotential height anomaly (in potential meters) and 700 hPa wind
 518 stream line averaged on the peak day (day 0), day 5 and lag day 10, composited for 85 positive
 519 NAO events (a, b and c) and for 82 negative NAO events (d, e and f). The brown lines represent
 520 the 1500 m isoheight. The dots represent the grid points where the 500 hPa geopotential
 521 height anomaly were significant at the 95% significance level.

522



523

524 Fig. 7. The 700 hPa moisture flux anomaly ($\text{g s}^{-1}\text{hPa}^{-1}\text{cm}^{-1}$, vectors) averaged from +5 day (5
 525 days after the peak day) to the +10 day and WWD frequency anomaly (percentage per day,
 526 shaded), composited for 85 positive NAO events (a) and for 82 negative NAO events (b). Panels
 527 (c) and (d) are the same as panels (a) and (b), respectively, but for precipitation anomaly (in 0.1
 528 mm d^{-1}). The brown lines represent the 1500 m isoheight and the red boxes represent the
 529 CAWTP region.



530

531 Fig. 8. Schematic diagrams showing how positive (a) and negative (b) NAO events influence
 532 the precipitation over the CAWTP region. The red and blue ellipses are the areas of the 500
 533 hPa positive and negative geopotential height anomalies. The blue lines represent the
 534 southern and northern branches of the 700 hPa westerly circulation. The black arrows
 535 represent the 700 hPa wind anomaly. The brown and gray dotted line represent troughs and
 536 ridges. The green and yellow areas represent the positive and negative precipitation
 537 anomalies.