Atmospheric rivers over the Bay of Bengal lead to northern Indian extreme rainfall

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1   **Atmospheric Rivers over the Bay of Bengal Lead to Northern Indian Extreme**

2   **Rainfall**

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Abstract

Atmospheric rivers (ARs), filamentary patterns of strong water vapor fluxes, play a

prominent role in global poleward moisture transport and have profound impacts on

extreme rainfalls (ERs). Previous AR research has mainly focused on the mid-latitude

regions, whereas the characteristics of ARs in low latitudes and their relationship with

local ERs remain largely unknown. This study investigates the spatiotemporal

characteristics of ARs over the Bay of Bengal and their relationship with ERs after

landing on the northern Indian subcontinent using the ERA-Interim reanalysis data.

During the study period from 1979 to 2011, a total of 149 ARs have been identified,

which feature a bimodal temporal pattern with more events observed in May and

October. The AR axes generally stretch northeastwards over the bay and land in

Bangladesh and Burma. 24% of ARs occurring during tropical cyclones implies a

possible connection between them, in addition to the similar intra-annual distribution. In

summer, as the tropical cyclones are weak and the northward water vapor flux decreases

due to topographic blocking of the Western Ghats, it is less likely to form intensified

water vapor pathway, though the atmospheric humidity is high in the study region.

Furthermore, a close correlation between ARs and ERs is manifested. A large proportion

of ARs would lead to ERs, with a small fraction of ERs occur after ARs. In addition,

although persistent ARs constitute the majority of identified events, rainfall intensity

will not be enhanced by the increase in AR duration. This study enriches the knowledge

of AR characteristics in low latitudes and provides new pathways to understand the

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41   hydrological cycles in the Indian Peninsula and the Bay of Bengal.

42   **Keywords:**

43   Atmospheric river; Bay of Bengal; Tropical cyclone; Extreme rainfall; ERA-Interim

44   dataset

1.   Introduction

45

46 Atmospheric rivers (ARs), featured by long, narrow corridors of enhanced water vapor

47 transport, are responsible for more than 90% of the horizontal water vapor transport

48 from tropics or subtropics into higher latitudes, and facilitate the precipitation that is

49 crucial to water supply (Zhu and Newell 1998; Ralph et al. 2011a; Bao et al. 2006).

50 Some strong ARs that contain large amount of water vapor associated with fierce winds

51 may lead to extreme rainfall (ER) and floods (e.g., Ralph et al. 2004; Lavers et al. 2011).

52 Consequently, ARs are widely recognized for the crucial role in the global water cycle

53 (e.g., Zhu and Newell 1998) and particularly those hydrometeorological extremes

54 (Neiman et al. 2008b).

55

56 In the mid-latitudes, these intensified water vapor plumes are often manifested as

57 moisture convergence in extratropical cyclone warm sectors and the pre-cold-frontal

58 low-level jet region (Ralph et al. 2004, 2006; Bao et al. 2006; Stohl et al. 2008). Two

59 moisture sources are considered responsible for AR formation: local moisture

60 convergence along the trailing cold fronts and direct poleward transport of tropical

61 moisture (Bao et al. 2006). In addition, Dacre et al. (2015) found that water vapor in the

62 cyclones' warm sector is more responsible for the generation of high water vapor

63 content, compared with long-distance transport of water vapor from subtropics, which

64 explains the high frequency of wintertime ARs in mid-latitudes. Furthermore, Zhu and

65 Newell (1998) speculated that tropical ARs may be related to large-scale convergence in

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66   boundary regions such as the intertropical convergence zone (ITCZ), rather than

67   baroclinic cyclones.

68

69   In the vicinity of ARs, cloud-top temperatures are colder, indicating that deeper clouds

70   (mid- and high clouds) are associated with deep convective systems that promote heavy

71   precipitation by strong upward motion up to the tropopause with a large amount of

72   cloud liquid (Ralph et al. 2004, 2011b). Most of the water vapor in ARs is transported

73   within the lowest 2.5 km of the atmosphere with moist-neutral stratification (Ralph et al.

74   2005). The combination of lower-troposphere moist neutrality and strong horizontal

75   winds creates an ideal avenue for strong orographic precipitation when ARs approach

76   high terrains (Ralph et al. 2011a; Lavers et al. 2011, 2012; Viale and Nunez 2010). For

77   example, ARs stretching from the eastern Pacific Ocean to western North America are

78   regarded as a primary meteorological contributor in flood generation and water

79   resources in California as ARs bring in 20-50% of the total precipitation in the long

80   term (Dettinger et al. 2011). Additionally, landfalling AR is found to be a strong

81   precursor to the ER in the western European seaboard: 8 of the 10 largest ER events

82   were preceded by ARs (Lavers and Villarini 2013b). The concept of ARs provides a new

83   and objective framework to examine and quantify atmospheric conditions related to ERs

84   (Ralph et al. 2006).

85

86   Previous techniques for AR identification are primarily based on the intensity and

5

87   spatial features of atmospheric water vapor (Gimeno et al. 2014). The intensity is

88   usually quantified by the vertically integrated water vapor (IWV) and vertically

89   integrated horizontal water vapor transport (IVT). IWV can either be retrieved from the

90   Special Sensor Microwave Imager (SSM/I) (Ralph et al. 2004), or calculated using

91   atmospheric reanalysis data; while IVT can be determined by using atmospheric

92   reanalysis data (Zhu and Newell 1998). Compared with IWV, IVT is more desirable for

93   AR detection due to its closer relationship with orographic precipitation (Neiman et al.

94   2002; Guan and Waliser 2015). In terms of geometric features, length, width and

95   length/width ratio are usually used in AR identification (Guan and Waliser 2015).

96   However, thresholds of these measures are not universal but determined region by

97   region (Gimeno et al. 2014). For instance, Neiman et al. (2008a) suggested that narrow

98   plumes of SSM/I with IWV values >20 mm (i.e., >2000 km long and <1000 km wide)

99   can be identified as ARs along the west coast of North America, whereas ARs at the

100   Midwestern USA are found to have higher IWV intensity (in the range of 30–55 mm)

101   and longer extension (about 3000 km) (Moore et al. 2012).

102

103   Although ARs in mid-latitudes are well studied by previous research, characteristics of

104   ARs in lower latitudes are remaining largely unknown. Meanwhile, the Bay of Bengal is

105   under strong influence of Indian monsoon, South Asian monsoon and notable tropical

106   cyclones and thus becomes vulnerable to devastating rainfall and floods that can be

107   induced by ARs. Additionally, inland penetration of ARs in concomitant with the

6

108   Himalayas may facilitate the topographic ERs in this region. As such, we focus on ARs

109   over the Bay of Bengal and examine their relationship with ERs over the Indian

110   subcontinent (Figure 1).

111

112   The objectives of the study are (1) to investigate the spatiotemporal characteristics of

113   ARs over the Bay of Bengal, and (2) to examine their relationship with ERs in the

114   northern Indian subcontinent. In the remainder of this paper, we first introduce an

115   algorithm for AR identification and apply it over the Bay of Bengal. Next, based on the

116   results of identified ARs, we analyze their characteristics and relationship with ERs in

117   the northern Indian subcontinent.

118

119   **2.   Methodology**

120   **2.1 Data**

121   The ERA-Interim dataset (Dee et al. 2011) at a 1.5° resolution spanning from 1979 to

122   2011 is used in this study. The ERA-Interim dataset features better representation of

123   hydrological cycle (Berrisford et al. 2011) and has been widely used in identifying

124   moisture origins (Zhao et al. 2015), delineating water vapor transport pathways (Dong

125   et al. 2016), etc. Especially, it can represent key AR characteristics (Guan and Waliser

126   2015). In the dataset, the specific humidity and the zonal and meridional wind fields are

127   available at a 6 h resolution, while the precipitation and evaporation are provided at a 3

128   h resolution. The specific humidity and horizontal wind fields are used in the AR

129   identification algorithm, while evaporation between 1000 hPa and 175 hPa is used to

130   correct precipitation by taking condensation as a part of precipitation. Also, as the

131   rainfall is available at a relative coarse spatial resolution (i.e., 1.5°), some

132   subsynoptic-scale gradients, such as those associated with orographic processes over

133   mountain barriers, may be smoothed out (Rutz et al. 2014). Nevertheless, the

134   ERA-Interim precipitation has the same spatial coverage and consistency between the

135   AR and ER results that will facilitate the exploration of their correlation in the later

136   sections.

137

138   2.2 ER Identification

139   An ER is defined as the 90th percentile of the maximum daily rainfall over the study

140   area during 1979-2011 (Champion et al. 2015). In addition, only rainy days with daily

141   precipitation larger than 1 mm are included (Eiras-Barca et al. 2016). The 90th percentile

142   threshold is calculated separately for each month (Figure 2).

143

144   2.3 AR Identification

145   We modify the method in Lavers and Villarini (2013a) for central United States in two

146   aspects: (1) Two reference latitudes are used in this study: 24°N and 27°N that represent

147   the coastal and interior areas respectively, in consideration of the variation of IVT

148   strength along latitude; (2) Equivalent width and direction thresholds are used in the

149   back trajectory algorithm in order to filter out invalid identification results.

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151   The detection method is based on the IVT field derived from specific humidity and

152   wind fields between 1000 and 300 hPa given by

153

$$\text{IVT} = \sqrt{\left(\frac{1}{g}\int_{1000}^{300} qu\,dp\right)^2 + \left(\frac{1}{g}\int_{1000}^{300} qv\,dp\right)^2} \qquad (1)$$

154   where q is the specific humidity in kg kg^{-1}, u and v are the zonal wind and meridional

155   wind in m s^{-1}, respectively, g is the acceleration of gravity in m s^{-1} (Neiman et al. 2008a;

156   Lavers and Villarini 2013a, 2013b).

157

158   Monthly IVT strength thresholds are applied in the AR identification considering the

159   notable intra-annual variability of IVT intensity (Champion et al. 2015) and ER

160   occurrence. The IVT threshold is determined as follows. The maximum values of IVT at

161   reference latitudes (i.e., 24°N and 27°N) at 1200 UTC of each day are first regrouped by

162   month, then the 85th percentiles of each group are determined as the monthly thresholds.

163   We note that IVT results smaller than 250 kg m^{-1} s^{-1} are excluded as they are too small

164   in winter (Rutz et al. 2014).

165

166   Besides the IVT strength threshold determination, the two reference latitudes are also

167   used in locating the ending point of AR axis (i.e., the central line of an AR). By

168   identifying ARs based on the three latitudes of the study area (i.e., 24 °N, 25.5 °N and

169   27 °N), we found the results based on 24°N and 27°N are sufficient to cover all AR

170   events. Thus, we employ 24°N and 27°N as reference latitudes (Full details of reference

171   latitude selection refer to Appendix).

172

173   The next step of AR identification is to determine the AR axis, which in combination

174   with geometry thresholds will be used to filter out the objects with less physical

175   meanings: as ARs refer to the strong poleward water vapor transport, objects not

176   satisfying such directionality need to be excluded. To start with, we define the ending

177   point of AR axis as the location of the maximum IVT at reference latitudes. Then a back

178   trajectory algorithm is used in the axis identification (Figure 3): we search for the

179   highest IVT by comparing the adjacent grid cells to the south, southeast and southwest,

180   which are selected because of the poleward transport feature, and such procedure is

181   iterated until a total of 14 successive grid points are obtained to form an axis. If 14 grids

182   are taken south of reference latitude, the identified axis would be more than 2000 km

183   long (Ralph et al. 2004).

184

185   Direction and width thresholds are then applied to distinguish ARs from monsoon and

186   tropical cyclones. Moisture convergence in a low-pressure center and intense surface

187   wind of summer monsoon are likely to be mistakenly identified as ARs if only IVT

188   strength and length are considered. The direction of an AR should satisfy the condition

189   that the included angle between mean transport direction and the line connecting two

190   endpoints of the axis is not greater than 45°. The mean transport direction is the mean

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191   moisture transport direction of grids along the axis. Here we use the equivalent width

192   (EW) to represent the width of an AR, which should be less than 1000 km (7 grids)

193   (Ralph et al. 2004; Neiman et al. 2008a). The EW is calculated as:

$$EW = \frac{area}{length} \tag{2}$$

194   where length is the number of the contiguous grids (14 in this case), area is the total

195   number of grids with IVT strength greater than the monthly threshold in the axis

196   coverage region, which is a sub-domain of the rectangular region extending diagonally

197   from (76.5°E，0°N) to (99°E，30°N): if the axis stretches zonally (meridionally), its

198   endpoints will determine the north-south (west-east) boundary of the coverage region.

199

200   To briefly summarize, AR identification includes three steps: (1) obtaining the original

201   AR occurrence time based on the IVT strength threshold; (2) conducting AR axis

202   identification and (3) filtering out the mistakenly identified AR structures with width

203   and direction thresholds. Figure 4 illustrates the brief workflow of AR identification

204   algorithm.

205

206   **3.   Results and discussion**

207   **3.1 IVT intensity**

208   Apparent seasonality in the atmospheric water vapor indicated by IVT is observed in the

209   study region (Figure 5). This pattern features high precipitation due to the strong

210   influence of southwesterly monsoon with abundant water vapor (Zhao et al. 2015). In

211   general, high IVT (≥ 250 kg m^{-1} s^{-1}) is observed in summer (JJA, Figure 5b) and autumn

212   (SON, Figure 5c) but winter (DJF, Figure 5d) and spring (MAM, Figure 5a). Annual

213   IVT exhibits a single-peak trend with the maximum in July as high as 600 kg m^{-1} s^{-1}. In

214   May (not shown) and autumn (SON, Figure 5c), the water vapor belt forms over the bay

215   and extends meridionally to the northeast of the study region. While in JJAS (Figure 5e),

216   the seasonal movement of the ITCZ, manifesting as monsoon, prompts the seasonal

217   rainfall variations. High rainfalls occur over the monsoon zone around 20° N stretching

218   northwestward from the head of the Bay of Bengal. And most of the summertime

219   monsoon rainfalls occur in association with the propagation of synoptic-scale

220   convective systems from the Bay of Bengal along the monsoon zone (Gadgil 2003;

221   Schneider et al. 2014). Also, as the East Asian monsoon becomes stronger in autumn,

222   water vapor brought by the southwesterly and East Asian monsoon converges near the

223   Longitudinal Range-Gorge Region in China (yellow box in Figure 5c). While the

224   wintertime IVT is relatively too small to impose apparent influence on the precipitation.

225

226   Inland penetration of ARs implies their possible impacts on areas that are distant from

227   the landfalling locations (Guan and Waliser 2015; Rutz et al. 2014). Considering the

228   high possibility of AR penetration, it is necessary to examine the possible extents of

229   ARs in the study region. As such, the dependence of IVT intensity on latitude is

230   assessed using their monthly thresholds of each latitude (Figure 6). A slight increase in

231   IVT is observed from 24°N to 25.5°N, which can be attributed to the high local

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232   evaporation from densely covered rivers in this area. Although IVT then decreases from

233   25.5°N to 27°N, the southern foot of the Himalayas, its relatively high value (larger than

234   250 kg m^{-1} s^{-1}) implies ARs may penetrate into the interior area of the study region.

235

236   **3.2 AR**

237   The intra-annual distribution of AR demonstrates a clear dual-peak pattern (peaking in

238   May and October) (Figure 6), which differs from single-peak pattern of ARs (peaking in

239   winter) observed in mid-latitudes. It is also noteworthy that the intra-annual variation in

240   AR does not follow that in IVT (single-peak pattern in summer), implying the

241   summertime decrease in AR events is caused not by the reduced IVT intensity but

242   possibly by the seasonal movement of water vapor belts: as the belt encounters the

243   Western Ghats (cf. Section 3.1), the high deficit in water vapor hampers the formation

244   of AR in summer.

245

246   To investigate the spatial pattern of ARs, AR axis is introduced as the indicator.

247   Comparable spatial occurrence patterns of ARs are observed in spring (Figure 7a) and

248   autumn (Figure 7c) with inland occurrence in Burma, Bangladesh and India. While less

249   occurrence of ARs is observed in summer, the intensity of AR is much stronger than

250   those in other seasons with deeper penetration (Figure 7b). The occurrence of AR is

251   significantly less in winter (Figure 7d). Figure 8a-8d show examples of ARs identified

252   in the four seasons.

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253

254   Considering the similar bimodal temporal distribution of ARs and tropical cyclones over

255   the Bay of Bengal (Li et al. 2012), we hypothesize that possible correlation may exist

256   between them. Using the tropical cyclone best-track data from the Joint Typhoon

257   Warning Center (JTWC), we compare the occurrence of ARs and tropical cyclones. The

258   results indicate that 24% of ARs occurred during tropical cyclones with most in the end

259   of the best track periods. We note that this percentage could be larger because ARs that

260   occurred after the tropical cyclones were excluded. Compared with the findings in

261   southeastern United States that only ~7% of ARs occurred during tropical and

262   subtropical cyclones (Debbage et al. 2017), it is indicated that ARs in the Bay of Bengal

263   are heavily influenced by tropical activity. Figure 8c shows the spatial pattern of a

264   typical AR that occurred during the active tropical cyclone period, indicating the role of

265   tropical cyclones in facilitating the longitudinal transport of water vapor that feeds ARs.

266

267   Based on the hypothesis, the low summertime occurrence of ARs may also be attributed

268   to other factors (e.g., vertical wind shear and vorticity (Li et al. 2012)) that limit the

269   formation of tropical cyclones, in addition to the weakened water vapor transport due to

270   topographic blocking (cf. Section 3.1). Also, water vapor belts are sometimes led by the

271   local convective systems to the East India during the summer monsoon (Figure 8e).

272

273   **3.3 Relevance to Extreme Precipitation**

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274　The relevance of ARs to ERs is investigated by testing the following hypotheses: 1) ERs

275　occur after ARs and 2) ARs lead to ERs. During the period from 1979 to 2011, a total of

276　1190 ERs have been identified with the spatial distribution concentrating mainly along

277　the south foot of the Himalayas (Figure 9), suggesting the importance of orographic

278　effect in the formation of ERs. Furthermore, we term those ERs that occur after ARs no

279　later than 24 hours and appear in the vicinity of the AR axis (i.e. within 1.5 degree from

280　the AR axis) as AR-related ERs (AERs) hereinafter. The above hypotheses are then

281　tested through two ratios: 1) r_{ER} (the occurrences of AERs over all ERs) and 2) r_{AR}

282　(the occurrences of AERs over all ARs). r_{ER} indicates the fraction of ERs that are

283　associated with ARs, whereas r_{AR} suggests the fraction of ARs that lead to ERs.

284

285　Noticeable seasonality is found in r_{ER} (Figure 10a) with higher values of ~14%

286　observed in October and November (the post-monsoon season) while the lowest values

287　in summer and winter. Such seasonality could be explained by the variability of

288　synoptic forcing and large-scale moisture availability (Eiras-Barca et al. 2016). During

289　the transitional monsoon season, AERs are favored by the abundant moisture conveyed

290　by ARs that satisfies the dynamic conditions for ERs. However, in the warm season,

291　precipitation tends to be more associated with convective systems (Lavers and Villarini

292　2013b; Champion et al. 2015; Eiras-Barca et al. 2016; Gadgil, 2003). Hence, the

293　moisture availability is no longer a restriction due to local moist convection favored by

294　higher temperatures. Meanwhile the tropical cyclones are less frequent, so ERs are

295   mostly attributable to local convective systems rather than ARs (Champion et al. 2015).

296

297   Apparent seasonality is found in r_{AR} as well (Figure 10b): wintertime values are much

298   higher than those in other seasons. Compared with r_{ER}, r_{AR} shows larger magnitude

299   over the whole year, indicating stronger correlation of AER with AR than with ER. In

300   particular, during the dry boreal winter, the high r_{AR} (i.e., >50%) highlights the crucial

301   role of ARs in forming ERs by supplying abundant moisture. As for the spatial

302   distribution of ERs, local topographic lift is found of great importance in facilitating

303   AERs. Therefore, higher winter half-year r_{AR} also indicates the potential of ARs in

304   inland penetration. On the other hand, the relatively low r_{AR} in spring could be

305   attributed to ARs that are not strong enough to penetrate deeply inland.

306

307   Based on the results of r_{ER} and r_{AR}, we thus infer that only a very small fraction of

308   ERs occur after ARs whereas a high portion of ARs can lead to ERs in the study region.

309   Furthermore, as the positive relationship is found between the duration of AR and the

310   rainfall/streamflow during AER (i.e. longer AR duration potentially induce higher

311   rainfall/streamflow, see Lavers and Villarini 2013a in central United States, Ralph et al.

312   2006 in northern California), we also examine such relationship in the study region.

313

314   ARs are categorized into three groups according to their durations: 1) AR-all: all AR

315   events are included; 2) AR-12h: AR events longer than 12 hours are included; 3)

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316   AR-18h: AR events longer than 18 hours are included. Results show that 70% of the

317   149 AR events last more than 12 hours, and up to 50% ARs fall into group AR-18h. It is

318   noteworthy that similar r_{AR} values of ~50% are found across the three groups (Figure

319   11), suggesting the AR duration has minimal influence on the intensity of ERs.

320

321   **4.   Concluding remarks**

322   In this study, we investigate the spatiotemporal characteristics of ARs over the Bay of

323   Bengal and their correlation with ERs in study area. In total, 149 ARs are identified

324   from 1979 to 2011 using a modified algorithm of AR identification in the IVT field,

325   over 50% of which last more than 18 hours. The modification includes width and

326   direction thresholds, in addition to length and IVT strength criteria, to filter out the

327   monsoon and tropical cyclones, which could be mistakenly identified as ARs in the

328   tropics. The choice of reference latitude is shown to be crucial in AR identification,

329   given the remarkable variation of IVT threshold along latitude. Two reference latitudes

330   are found to be sufficient to capture all AR events of the study region.

331

332   The intra-annual variation of AR occurrence shows a bimodal feature (i.e., peaking in

333   May and October) that is consistent with the temporal distribution of tropical cyclones

334   over the Bay of Bengal. These moisture corridors form over the bay, extend

335   northeastwards, and penetrate inland into Burma, Bangladesh and India. The

336   spatiotemporal characteristics of ARs demonstrate that the southwesterly moisture

17

337   transport brings abundant water vapor from the tropics to the northern continent and

338   indicate that tropical cyclones may enhance this poleward moisture transport. Moreover,

339   the high occurrence frequency of ARs in May is different with tropical cyclones but

340   similar with supercyclones (Li et al. 2012), implying that the intensity of tropical

341   cyclones might influence the occurrence of ARs. The low frequency of ARs in summer

342   could be explained by the lack of cyclones and the weak northward water vapor

343   transport due to topographic blocking.

344

345   Furthermore, a high correlation between ARs and ERs is verified. Although the fraction

346   of ERs occurring after ARs is relatively small, a large proportion of ARs lead to ERs. In

347   particular, the moisture conveyed by ARs plays an important role in forming ERs during

348   the dry boreal winter. Besides, local convective systems responsible for most summer

349   ERs explain the low occurrence of ERs induced by ARs in summer. On the other hand,

350   the fraction of ARs leading to ERs indicates the potential of ARs in facilitating the

351   landing of ERs considering the crucial role of mountains in the ER formation. In

352   addition, the duration of ARs has minimal influence on the intensity of ERs.

353

354   This study for the first time investigates the spatiotemporal characteristics of ARs in low

355   latitudes and their relationship with ERs. Although both are closely correlated with ERs,

356   the dual-peak pattern in occurrence and strong inland penetration of ARs in low

357   latitudes highlight their distinct features from ARs in mid-latitude. These may help

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358   better understand the weather extremes and hydrological cycles in this region. One of

359   the limitations that need to appreciate is the use of a single reanalysis dataset: analysis

360   based on more datasets can help generalize the findings of this study. Additionally, the

361   mechanism behind which tropical activities govern ARs needs further exploration.

362 **Acknowledgement**

368   **Appendix: ARs identified based on two reference latitudes**

369   In this work, we apply two reference latitudes in AR identification and obtain two

370   different groups of ARs. The two reference latitudes are capable to envelope the

371   variation of grid IVT strength over the continent where ARs land (Figure A1).

372   Furthermore, compared with the global AR dataset of Guan and Waliser (2015),

373   although the total number of identified ARs in our work (the oceanic area of a

374   rectangular region extending diagonally from (76.5°E，0°N) to (99°E，24°N)) during

375   1979-2011 is relatively smaller, more than half of our results overlap with the global

376   dataset. Both results show the high frequency of ARs in warm season and a reduction in

377   July (Figure A2). The disparity in the amount of ARs identified can be attributed to

378   three reasons: 1) exclusion of eastward and westward tracks in the back trajectory

379   algorithm to focus on the water vapor directly from the Bay rather than from the

380   Arabian Sea and the South China Sea across land; 2) a much higher fixed lower limit

381   (i.e. 250 kg m^{-1} s^{-1}) (Rutz et al. 2014); 3) exclusion of invalid ARs using equivalent

382   width and direction thresholds during monsoon season (JJAS).

383

384   In addition, the difference between the 24°N and 27°N AR groups might be due to three

385   possible causes: 1) transport delay, 2) reduction in IVT threshold and 3) geometry

386   thresholds related uncertainty. Geometry filtering acting on AR axis structures with

387   ending points on different reference latitudes is likely to yield different results, even if at

388   the same occurrence time. However, as this impact is difficult to quantify, we do not

389   include the assessment in this section.

390

391   To assess the differences caused by transport delay, we choose an average lag of 5 d

392   through sensitivity analysis, considering the time scale of moisture recycling ranges

393   from 2 d to 10 d over the study region and the fast feedback is observed in the summer

394   (van der Ent and Savenije 2011). Only those ARs in 24°N and 27°N occurring within

395   the average lag (i.e., 5 d) are considered to be successive events, by which 24% of the

396   AR events that identified at 27°N can be categorized as successive ones.

397

398   Then, we analyze the impact of IVT threshold reduction on the AR identification. Given

399   the lower IVT threshold at 27°N compared with 24°N, we thus apply the threshold at

400   27°N to the dataset of 24°N to examine the impact of IVT threshold reduction. Based on

401   the IVT threshold at 27°N, 185 ARs are identified at 24°N, of which 26 events overlap

402   with those identified at 27°N.

403

404   The different groups of ARs identified based on the two latitudes have a marginal

405   overlap of 6 events. Besides transport delay and IVT threshold reduction, there are still

406   37% mismatched events in the AR group based on 27°N, which could be partly

407   attributed to geometry related filtering.

408

409   Moreover, apparent differences are found between the identified AR groups based on

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410   different reference latitudes (Figure A3): 24°N, 27°N and the complete list (i.e., the AR

411   group consisting results from both 24°N and 27°N). Large fluctuations are observed in

412   the inter-annual AR occurrence frequency. For instance, the annual AR occurrence

413   ranges between zero and fourteen in the complete list, with the year 1988 witnessing the

414   most AR events. Also, it is noteworthy that bursts of AR occurrence are observed in the

415   24°N group during each decade, implying the existence of inherent decadal frequency

416   of ARs in this region. On the contrary, the occurrence fluctuation of 27°N groups is

417   relatively milder in the study period with slightly growing trend.

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Atmospheric rivers over the Bay of Bengal

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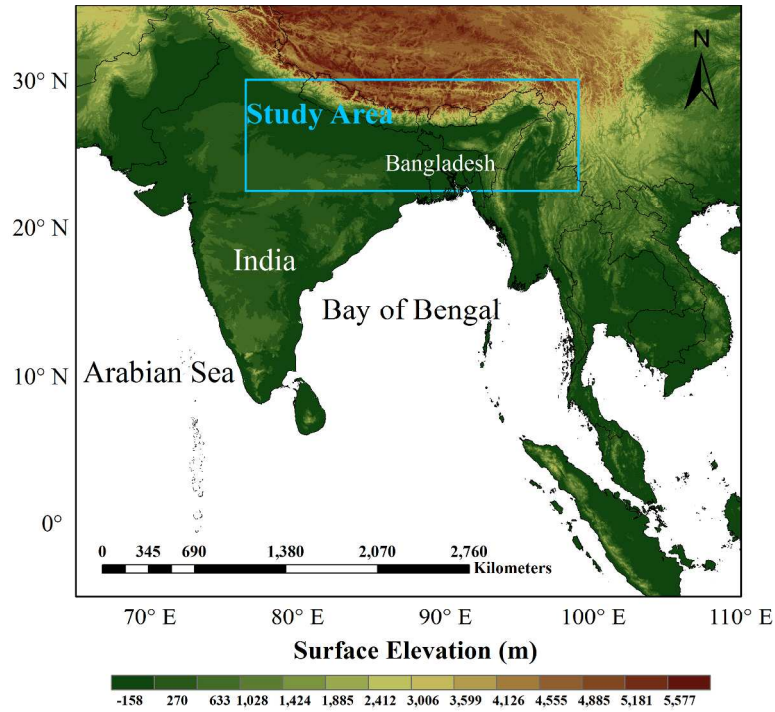
Figure 1 Observed (30 arc s) topography (shaded in m) for the Bay of Bengal (BOB) and part of Asia. The location of study area is shown in blue box (from (76.5°E, 22.5°N) to (99°E, 30°N)). The west-east boundary is based on the extent of BOB and the south-north boundary represents the northern continental area between coast and the Himalayas. The country boundaries are represented by black lines.

297x210mm (300 x 300 DPI)
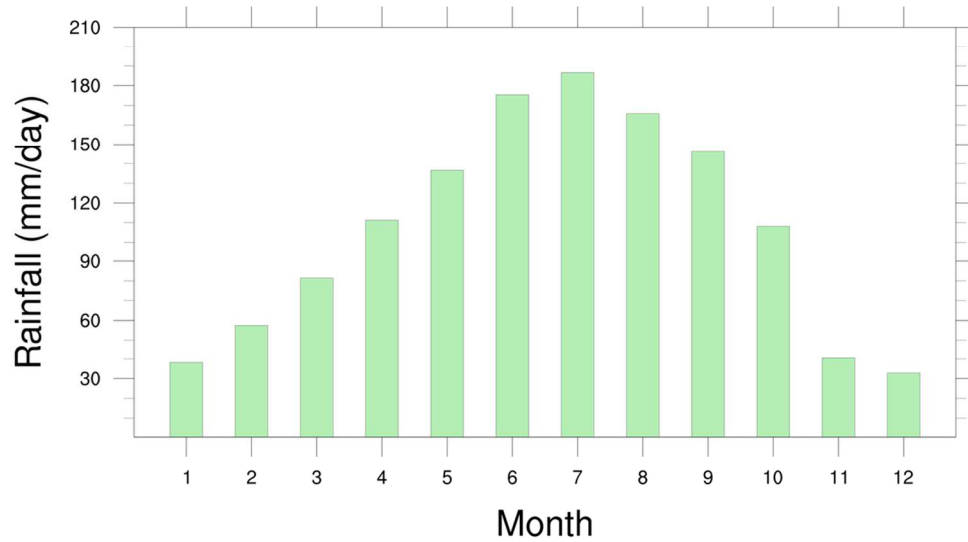
Figure 2 Monthly extreme rainfall thresholds defined as the 90th percentile of the maximum daily rainfall (larger than 1 mm) over the study region during 1979-2011.

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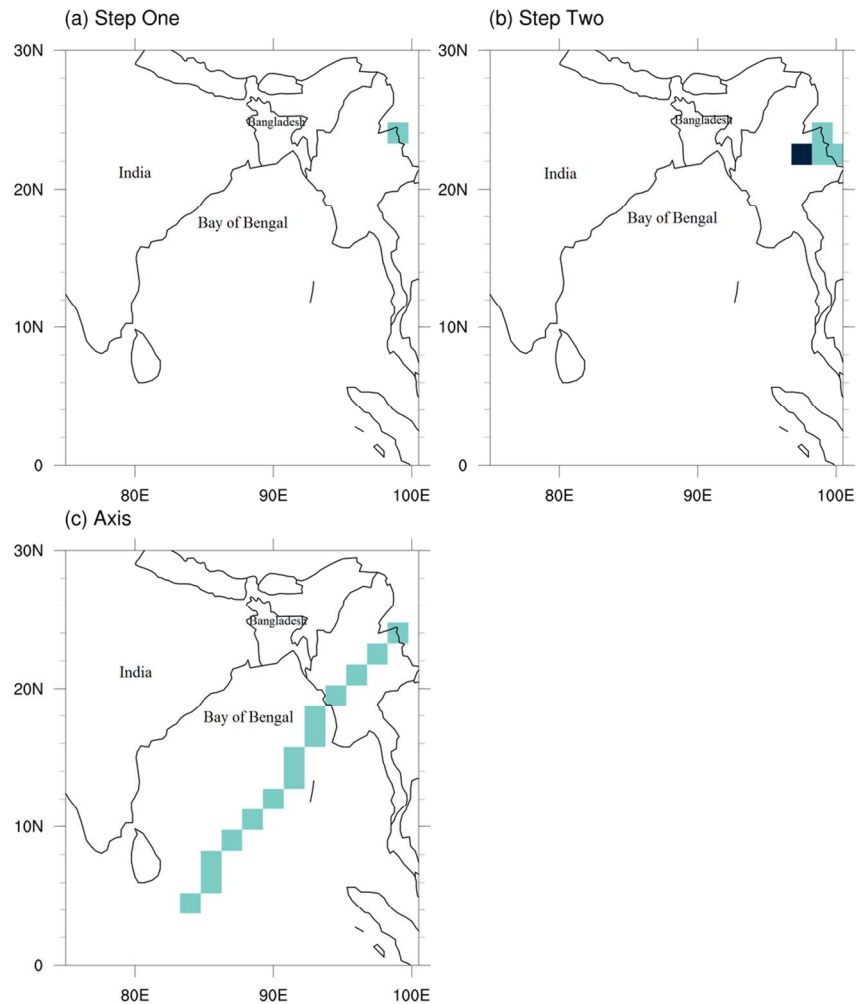
Figure 3 The AR axis structure identification algorithm. (a) The blue grid indicates the location of maximum IVT between 76.5°E and 99°E at 24°N (reference latitude), which serves as the ending point of the axis. (b) The next step is to search the adjacent grid cells (to the south/ southeast/ southwest) for highest IVT. The black grid represents the selected highest IVT location. (c) Repeat step two for 13 times and obtain the axis structure (the connected blue line).

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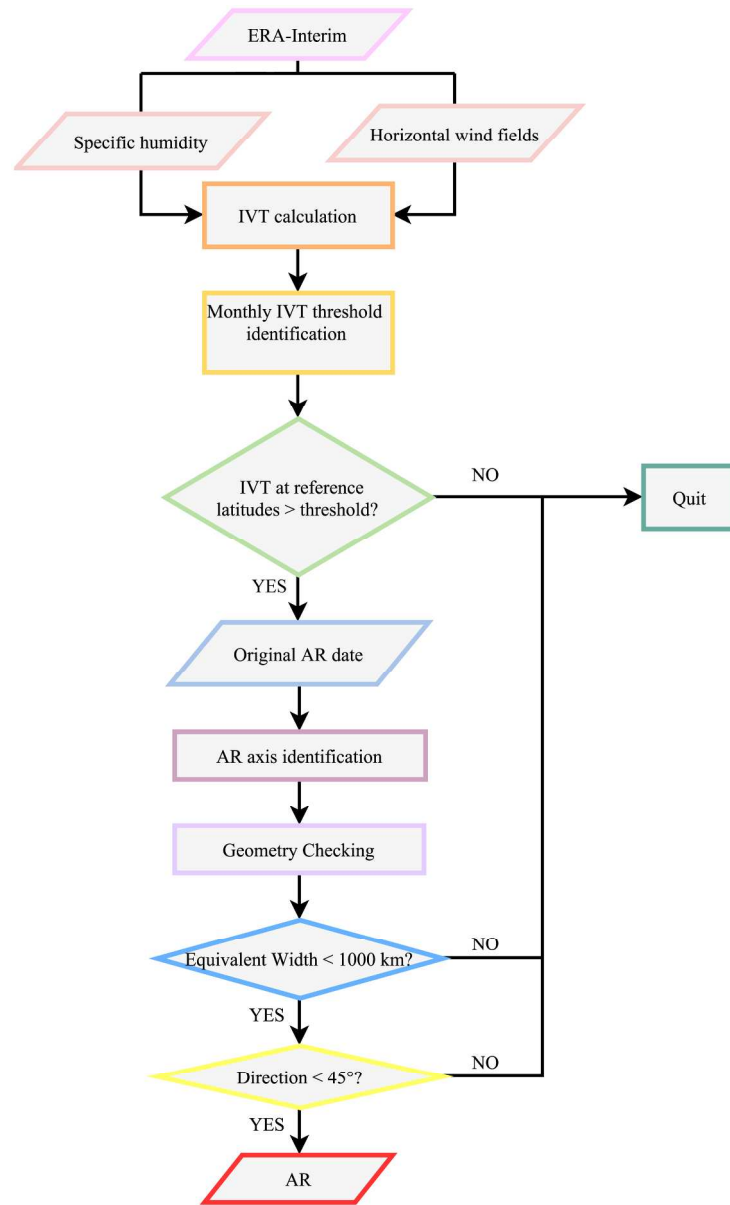
Figure 4 The brief framework for AR identification algorithm.

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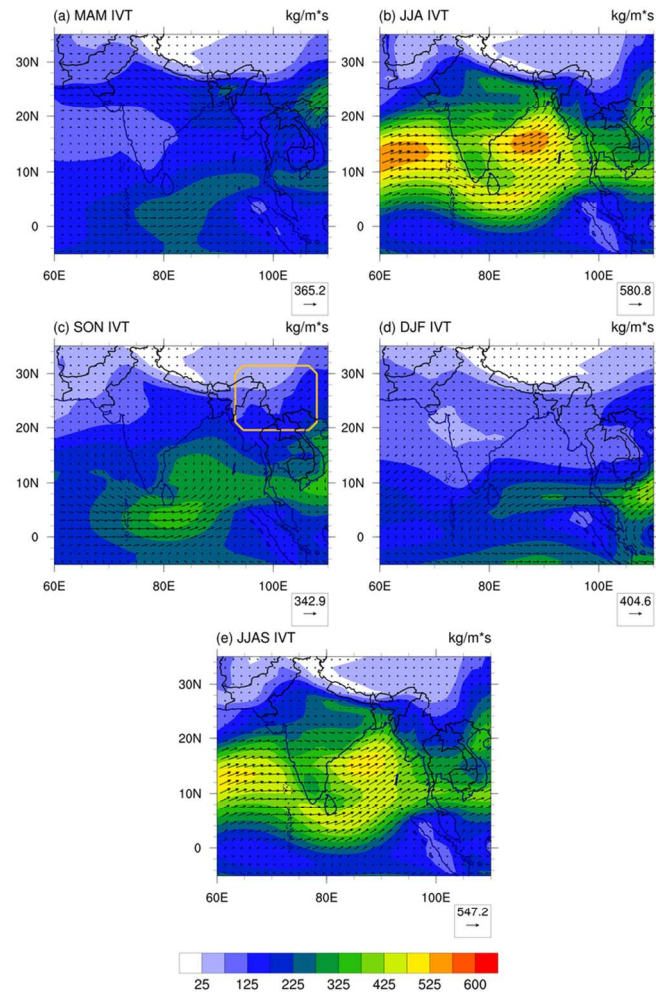
Figure 5 The seasonal mean IVT (in kg/m^{-1}s^{-1}) in (a) spring, (b) summer, (c) autumn, (d) winter and (e) monsoon season. The yellow box (from (21°N, 95°E) to (30° N, 106°E)) represents the spatial scope of Longitudinal Range-Gorge Region (Wu et al. 2012).

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Figure 6 The monthly IVT strength thresholds at 24°N (blue), 25.5°N (red) and 27°N (green) in the study area (left axis). The IVT strength threshold is determined as follows. The maximum values of IVT at certain latitudes at 1200 UTC of each day are first regrouped by month, then the 85th percentiles of each group are determined as the monthly thresholds. The purple line of 250 kg m^{-1} s^{-1} represents the lower limit of IVT strength (left axis). The chocolate line displays the intra-annual distribution of ARs identified over the Bay of Bengal during 1979-2011(right axis).

Figure 7 The spatial occurrence of ARs that is shown as the seasonal-accumulative occurrence frequency of AR axis in each grid.

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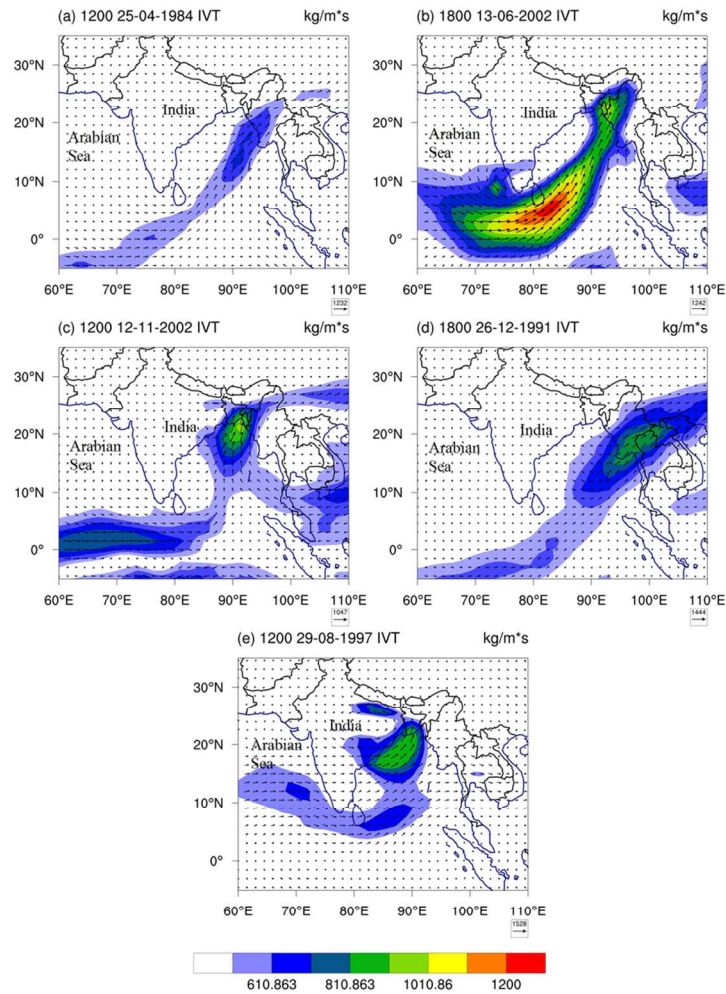
Figure 8 Examples of five ARs shown in IVT (in kg m^{-1}s^{-1}) at (a) 1200UTC 25th April 1984, (b) 1800UTC 13th June 2002, (c) 1200UTC 12th November 2002, (d) 1800UTC 26th December 1991, and (e) 1200 UTC 29th August 1997. Note that the lower limit is the IVT threshold of the corresponding month.
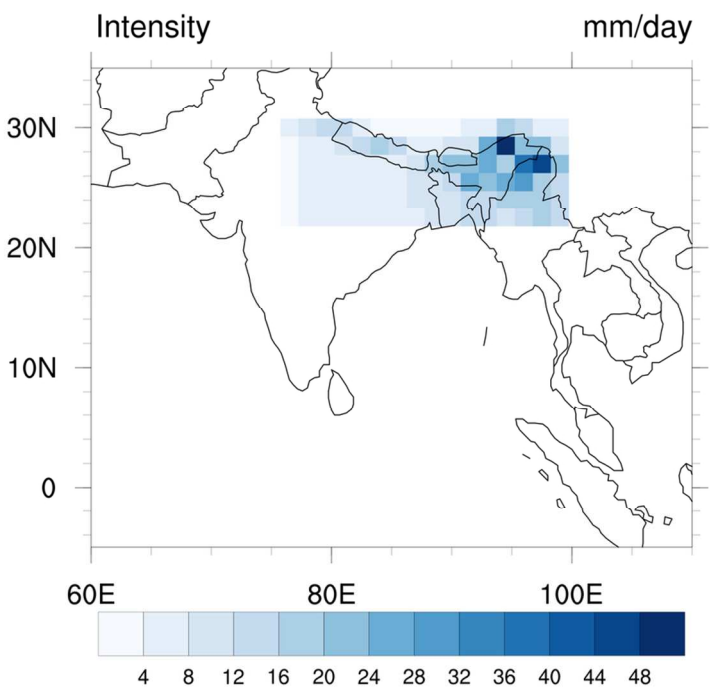
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Figure 9 the spatial distribution of mean ER intensity over study area.

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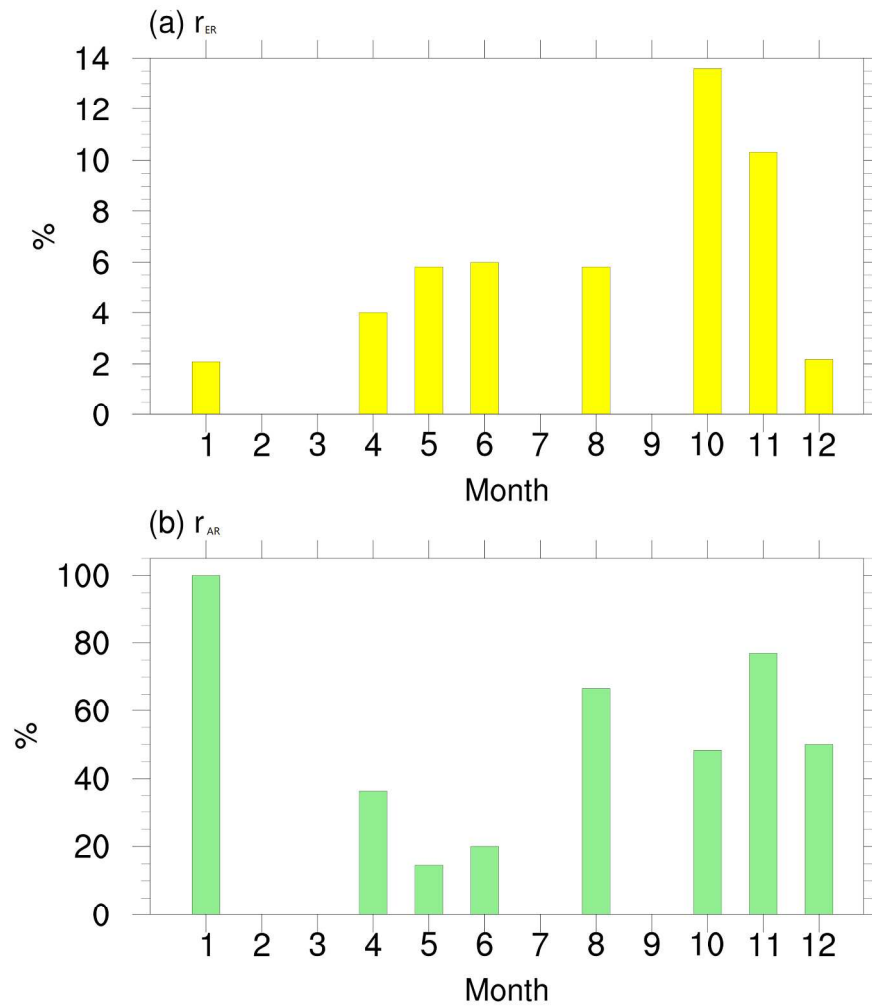
Figure 10 (a) and (b) are r_{ER} (the occurrences of AREs over all ERs) and r_{AR}(the occurrences of AERs over all ARs) respectively.
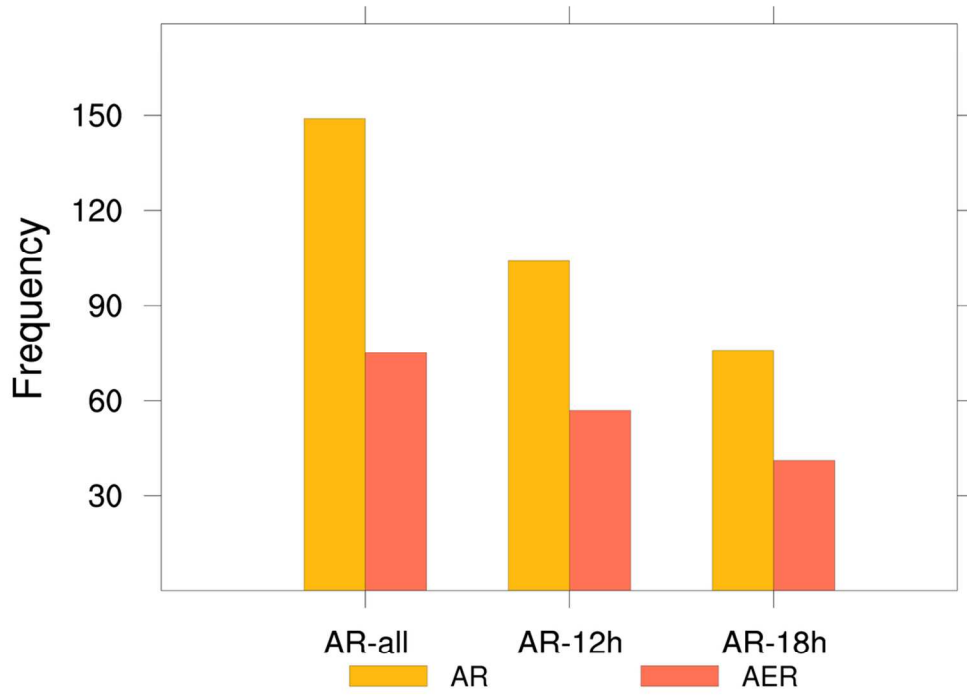
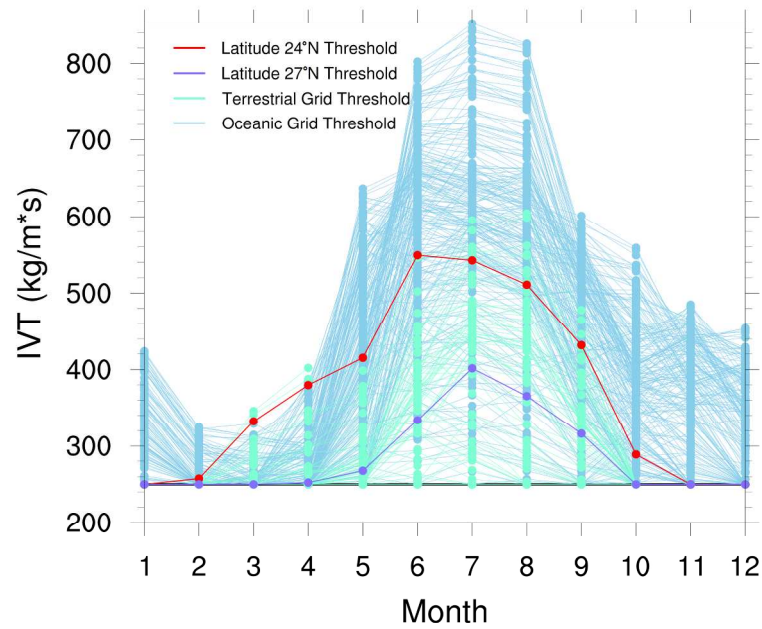Figure 11 Frequency of AERs in three groups.

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Figure A1 Monthly variations of IVT strength threshold of different areas: oceanic grids (blue, from (76.5°E, 0°N) to (99°E, 22.5°N)), terrestrial grids (green, from (76.5°E, 24°N) to (99°E, 30°N)), reference latitude 24°N (red), and reference latitude 27°N (purple). The black line indicates the lower limit of IVT threshold of 250 kg m^{-1} s^{-1}.
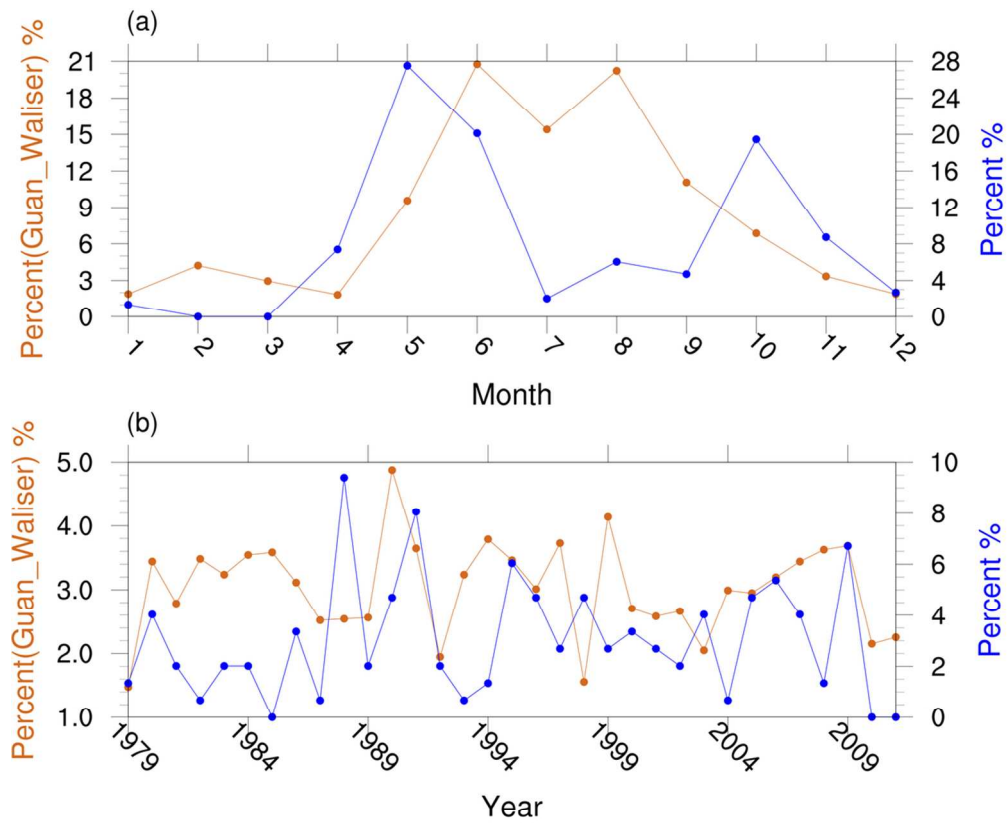
Figure A2 The (a) intra-annual and (b) inter-annual AR distribution of Guan and Waliser global AR dataset (brown) and our results (blue).
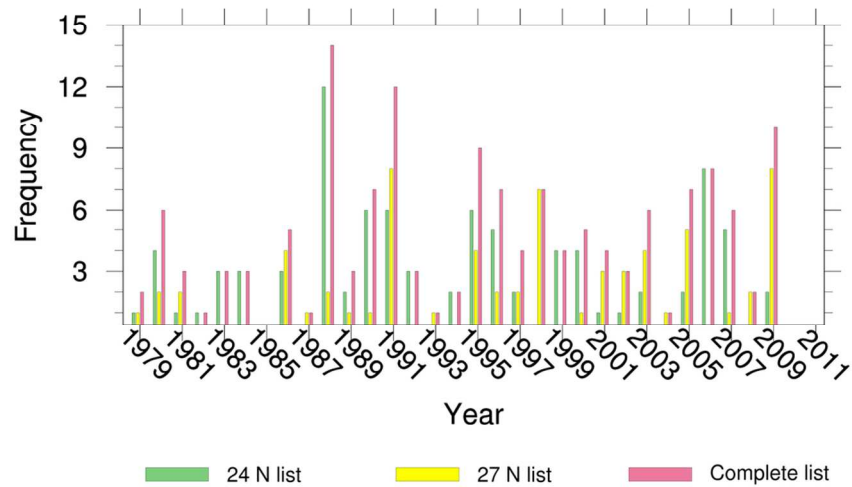
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Figure A3 Time series of identified ARs over 1979-2011. The green bars and yellow bars show the 24°N and 27°N list respectively, and the pink bars represents the complete list.

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Figure Captions:

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