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The Imprint of Strong-Storm Tracks on Winter Weather in North America

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Abstract

1

2   Northern Hemisphere winter storm tracks and their relation to winter weather are investi-

3 gated using CFSR data. Storm tracks are described by isentropic PV maxima within a Lagran-

4 gian framework; these correspond well with those described in previous studies. Our diagnostics

5 focus on strong-storm tracks, which are comprised of storms that achieve a maximum PV ex-

6 ceeding the mean value by one standard deviation. Large increases in diabatic heating related to

7 deep convection occur where the storm tracks are most intense. The cyclogenesis pattern shows

8 that strong storms generally develop on the upstream sectors of the tracks. Intensification hap-

9 pens towards the eastern North Pacific and all across the North Atlantic Ocean, where enhanced

10 storm track-related weather is found.

11   In this study, the relation of storm tracks to near-surface winds and precipitation is evalu-

12 ated. The largest increases in storm track-related winds are found where strong storms tend to de-

13 velop and intensify, while storm precipitation is enhanced in areas where the storm tracks have

14 the highest intensity. Strong storms represent about 16% of all storms but contribute 30-50% of

15 the storm precipitation in the storm track regions. Both strong-storm related winds and precipita-

16 tion are prone to cause storm-related losses in the eastern US and North American coasts. Over

17 the oceans, maritime operations are expected to be most vulnerable to damage offshore of the US

18 coasts. Despite making up a small fraction of all storms, the strong-storm tracks have a signifi-

19 cant imprint on winter weather in North America potentially leading to structural and economic

20 loss.

1.   INTRODUCTION

Two well-documented mid-latitude winter storm tracks in the Northern Hemisphere (NH) affect North American weather and climate: the Pacific storm track which extends eastward across the North Pacific Ocean, and the Atlantic storm track which extends northeastward across the North Atlantic Ocean. Elsewhere in the NH mid-latitudes there is the Mediterranean storm track which spans eastward across the Mediterranean Sea to the Middle East (Hoskins and Hodges 2002). The storm tracks are characterized as large narrow bands of high baroclinic instability along which individual storms tend to propagate, and are maintained by the continuous downstream development of baroclinic disturbances (Simmons and Hoskins 1979; Wallace et al. 1988; Orlanski and Chang 1993). The upper-tropospheric winds (i.e., the 200-hPa jet stream) and divergence aloft produce cyclonic circulation poleward of the zonal flow, enhancing cyclonic shear and generating upstream confluence that can predominantly maintain the mean baroclinicity needed for continued downstream eddy activity (Wallace et al. 1988; Hoskins and Valdes 1990). Even in cases of weaker instability, the downstream radiation of kinetic energy in the form of ageostrophic geopotential fluxes contributes to the growth and intensification of new eddies at the expense of upstream decaying eddies (Simmons and Hoskins 1979; Orlanski and Chang 1993). Baroclinic disturbances propagate downstream as large-scale wave packets with a group velocity that primarily dictates the speed at which new eddies develop (Orlanski and Chang 1993).

Many factors influence the NH storm track distribution, including sea surface temperature (SST) gradients, uneven heating, and orography (Hoskins and Valdes 1990; Held 1993;

1

Brayshaw et al. 2008, 2009; Chang 2009). A strong mid-latitude SST gradient alongside a reduced subtropical SST gradient will generally strengthen the storm tracks and shift them poleward (Brayshaw et al. 2008). In the North Atlantic the large SST gradient formed by the protrusion of the warm Gulf Stream into the cool higher latitude ocean induces surface wind convergence on the warm side of the Gulf Stream front, intensifying the vertical wind velocity and vertical instability, in turn enhancing convection and storm development (Minobe et al. 2008, 2010).

Uneven diabatic heating induced in part by land-sea temperature contrasts plays a role in storm track modulation (Hoskins and Valdes 1990; Chang 2009). As cool westerly flow off the land passes over warmer western oceans, the surface air warms rapidly, triggering the generation of surface sensible heat fluxes that act to destabilize the atmosphere (Mak 1998). The sensible heat fluxes counter the damping effect of poleward eddy heat fluxes, preserving baroclinicity at the surface and maintaining the storm tracks through the development of unstable waves aloft (Hotta and Nakamura 2011). Asymmetries in diabatic heating partly account for the greater strength of the Atlantic storm track compared to the Pacific storm track, despite the lower baroclinicity in the Atlantic (Chang 2009). For instance, the large land-sea temperature gradient in winter induced by strong air mass contrasts between cold air over northeastern North America and warmer air over the Gulf Stream form a region of particularly high baroclinic instability along an axis that follows the North American east coast (Brayshaw et al. 2009). Storms tend to deepen and intensify leeward of the Appalachian Mountains (Colucci 1976), and the baroclinic zone over the North American east coast promotes the further amplification of storms, including nor'easters (Davis and Dolan 1993). Additionally, the strength and areal width of marine storms are determined by the intensity of the diabatic heating (Mak 1998).

44   As for orographic influences, mountainous terrain mainly acts to suppress storm track ac-

45   tivity by blocking or deflecting the westerly flow over land (Chang 2009). The Rocky Mountains

46   deflect westerly Pacific cyclones/storms southward which leads to a southwest-northeast (SW-

47   NE) tilt in the upper tropospheric jet, the subsequent downstream flow, and the Atlantic storm

48   track, dynamically separating the Northern Hemisphere storm tracks (Brayshaw et al. 2009;

49   Chang 2009). The Atlantic track lies coincident with the SW-NE axis of the low-level baroclinic

50   zone that follows the North American east coast, further enhancing cyclonic activity in the re-

51   gion of the Atlantic storm track (Brayshaw et al. 2009).

52   During winter, the Pacific and Atlantic storm tracks are collocated with climatological

53   precipitation maxima that exceed 6 mm day^{-1} (Adler et al. 2003; Hawcroft et al. 2015; Xie et al.

54   2017). Extremely high precipitation is produced primarily by extratropical storms with the most

55   heavily precipitating storms contributing substantially to the winter climatological precipitation

56   (Maddox et al. 1979; Hawcroft et al. 2012; Pfahl and Wernli 2012). In general in the NH, over

57   half of the mean total winter precipitation in the mid-latitudes is associated with frontal systems

58   and related cyclonic activity (Catto et al. 2012). Specifically in North America, over 70% of win-

59   ter precipitation is associated with low-level cyclonic activity (Hawcroft et al. 2012). It has also

60   been found that precipitation and upper-level zonal flow are highly correlated over the mid-lati-

61   tude oceans and over land upstream of high orography, supporting the notion that strong baro-

62   clinic cyclones aloft lead to large accumulations of precipitation at the surface (Maddox et al.

63   1979; Garreaud 2007; Pfahl and Wernli 2012). Accordingly, storm track modulation can be asso-

64   ciated with changes in the frequency of extreme precipitation and wind events, which can pro-

65   foundly affect a region's climate (Chang et al. 2002; Ma and Chang 2017). This can happen if a

66   northward shift and deepening of the semi-permanent Aleutian Low in the high latitudes of the

67 North Pacific Ocean occurs as it can then draw the Pacific storm track poleward and subse-

68 quently amplify winter precipitation in northwestern North America (Salathé 2006).

69   Previous studies have used different variables and metrics to represent storm tracks, in-

70 cluding mean sea level pressure (MSLP), geopotential height, and the meridional component of

71 the upper tropospheric wind (e.g., Gulev et al. 2001; Hoskins and Hodges 2002; Raible 2007).

72 MSLP and 500-hPa geopotential height are dominated by large scales, making small-scale, high-

73 frequency features like cyclones difficult to identify without bias toward larger, slower disturb-

74 ances (Wallace et al. 1988; Hoskins and Hodges 2002). The upper-level meridional wind tends to

75 better capture the higher frequencies and reveals downstream-developing wave trains along the

76 storm tracks (Chang and Orlanski 1993; Berbery and Vera 1996). Low-level relative vorticity

77 and isentropic potential vorticity (PV) are also useful to track storms because of their dependence

78 on higher order derivatives that allows for the detection of small-scale features such as cyclogen-

79 esis and cyclolysis (Hoskins and Hodges 2002). PV, in particular, is an ideal dynamical tracer

80 because of its conservation properties in an adiabatic, frictionless flow (Holton 2004). In the

81 Northern Hemisphere, a positive (cyclonic) PV anomaly, which generally corresponds to an up-

82 per-tropospheric pressure trough, induces a vortex with cyclonic circulation (Hoskins et al. 1985;

83 Hoskins and Hodges 2002). Because PV considers both absolute vorticity and static stability, it

84 encapsulates many of the dynamic and thermodynamic properties of the atmospheric circulation

85 while also conforming to the principle of invertibility, which establishes that the 3-dimensional

86 wind and temperature fields are induced by the PV structure if relatively fast-moving waves are

87 neglected (Hoskins et al. 1985; Hoskins 1997).

88   This study discusses the characteristics of the storm tracks as constituted by storms that

89   achieve high potential vorticity and will thus be called "strong-storm tracks". The primary objec-

90   tives of the study address the following questions: (1) how do strong-storm tracks relate to sur-

91   face weather and diabatic heating distributions?, and (2) what are the potential damaging effects

92   of very high near-surface winds and precipitation rates associated with the strong-storm tracks

93   that could lead to structural and economic loss in North America? We also discuss the robustness

94   of the results by using an independent dataset of observed precipitation.

95   The structure of the article is as follows: Section 2 describes the datasets and cyclone

96   tracking method used. Section 3 discusses the properties of the strong-storm tracks that affect

97   North America's winter weather, while Section 4 examines the relation between the strong-storm

98   tracks and the potential destructive effects of the associated wind and precipitation. Section 5

99   summarizes the key findings.

100

101   2.   DATA AND METHODOLOGY

102   *a.   Datasets*

103   The National Centers for Environmental Prediction (NCEP) Climate Forecast System Re-

104   analysis (CFSR; Saha et al. 2010) product is the most recent complete compilation of global rea-

105   nalysis data generated by NCEP. The CFSR couples the atmosphere, ocean, land surface, and sea

106   ice to provide our best 4D view of the Earth's natural state, constrained by observations, every 6

107   hours. The global atmospheric data have a horizontal grid spacing of 38 km, 64 vertical sigma-

108   pressure levels and are archived on a $0.5° \times 0.5°$ latitude-longitude grid. The gridded statistical

109   interpolation (GSI) scheme assimilates atmospheric variables including global precipitation rates

110   derived from rain gauge and satellite observations into CFSR. The coupled assimilation provides

111   a more complete and better-quality dataset of precipitation than past NCEP reanalyses that ne-

112   glect coupling in the data assimilation, with better correspondence between the model physics

113   and observed precipitation (Saha et al. 2010; Wang et al. 2011). The accuracy of CFSR is im-

114   proved over past NCEP reanalyses in part because of higher spatial and temporal resolutions, the

115   assimilation of bias corrected observations, and the coupling to sea ice and the ocean (Saha et al.

116   2010). Despite the improvements in CFSR, an artificial discontinuity around October 1998 in the

117   wind and precipitation time series has previously been found. The ingestion of satellite observa-

118   tions often marks the onset of artificial trends in other reanalysis datasets (Bengtsson et al. 2004),

119   and CFSR is no different. The discontinuity in CFSR is thought to be due to the introduction of

120   the assimilation of data from the low-earth polar-orbiting Advanced TIROS Operational Vertical

121   Sounder (ATOVS) satellite, which contributes to less spin-up of the initial moisture, resulting in

122   a more humid atmosphere after 1998 (Saha et al. 2010; Chelliah et al. 2011; Wang et al. 2011;

123   Zhang et al. 2012). Our study uses CFSR data for DJF from 1980-2010 and examines this dis-

124   continuity to assess the potential effects on our results.

125   The Global Precipitation Climatology Project (GPCP; Huffman et al. 2009) 1-degree

126   daily dataset of precipitation accumulations centered on 12:00 UTC is used to complement the

127   reanalysis information. Since October 1996, the GPCP has provided high quality, high resolution

128   global precipitation data. The daily dataset is derived from the GPCP Version-2 Monthly Precipi-

129   tation Analysis by combining in situ data (i.e., surface rain gauges) with histograms of 3-hourly

130   infrared brightness temperatures from geosynchronous-orbit satellite infrared data and precipita-

131   tion derived from atmospheric parameters retrieved from low earth orbit satellites (Huffman et

132   al. 2001; Adler et al. 2003; Pendergrass 2015). In this study, GPCP daily precipitation is used for

133   DJF from 1999-2010 and is taken as "ground truth" although some estimates, particularly over

134   oceans, may be less reliable (Adler et al. 2012).

135

136   *b.   Tracking of Storms*

137   Small scale isentropic potential vorticity maxima at the 320 K level (hereafter PV_{320}) at

138   6-hourly intervals are objectively tracked in DJF for 1980-2010 following the Lagrangian ap-

139   proach discussed in Hoskins and Hodges (2002). The method first identifies cyclones as PV_{320}

140   anomalies that exceed 0.5 Potential Vorticity Units (PVU, where 1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1}) on

141   a NH polar stereographic projection, which helps to prevent latitudinal bias in the identification

142   of cyclones at high latitudes (Sinclair 1997). The PV_{320} threshold of 0.5 PVU is significantly low

143   to account for most possible storms: in this case about 296 cyclones per DJF season are identi-

144   fied that satisfy the post tracking filters (discussed below). The 320 K isentrope is chosen as the

145   level of analysis as it resides in the mid-upper troposphere near the upper-level jet stream (Fig. 1)

146   where Rossby wave-induced baroclinic instability tends to occur (Hoskins 1991). The PV_{320}

147   anomalies are produced by applying a spherical harmonic analysis to the PV_{320} field and remov-

148   ing the background planetary scale waves with total wavenumbers less than or equal to 5 and re-

149   ducing the resolution to T42 to reduce noise. Additionally, a spectral taper is applied to the spec-

150   tral coefficients to further reduce noise (Hoskins and Hodges 2002). This has been found to be a

151   conservative but useful approach when examining fields that are dominated by a large scale

152   background and are very noisy at high resolutions and focuses on the synoptic spatial scales of

153   cyclones. The identified PV_{320} maxima are initially linked using a nearest neighbor method to

154   form tracks and are then refined using a constrained optimization approach which swaps points

155   between tracks to maximize the track smoothness (Hodges 1994, 1995). Constraints are applied

156   adaptively for maximum propagation speed and track smoothness (Hodges 1999) suitably chosen

157   for the extra-tropics.

158   Following completion of the tracking, a filter is applied to retain only those cyclones that

159   last at least 2 days and travel farther than 1000 km. These conditions act as spatial and temporal

160   filters to remove short duration or semi-stationary eddies. Considering that extratropical storms

161   at 320 K in the NH have an average PV_{max} = 3 PVU and a standard deviation $(PV_{max})_{SD}$ = 1.3

162   PVU, we define "all-storm tracks" as those shaped by storms with maximum PV that exceed a

163   low threshold of $\overline{PV}_{max} - 1.5 \times (PV_{max})_{SD} \cong 1$ PVU. As apparent in Fig. 2, this threshold cap-

164   tures weak cyclogenesis and provides a large number of cases for the analysis: on average, about

165   259 storms per season that satisfy the post tracking filters comprise the extratropical NH all-

166   storm tracks.

167   Storms in the Pacific and Atlantic Oceans have an average PV_{max} of 3.8 PVU with a

168   standard deviation of 1 PVU (Both regions have the same values, despite being computed sepa-

169   rately). "Strong-storm tracks" represent those storms in the Pacific and Atlantic Oceans with

170   maximum PV that exceeds a higher threshold of $\overline{PV}_{max} + 1 \times (PV_{max})_{SD}$ = 3.8 PVU +

171   1 PVU = 4.8 PVU as also noted in Fig. 2. Strong storms represent about 16% of all storms that

172   develop in both regions and correspond to similar percentiles of the storm strength distribution in

173   each basin. On average, 9 (6) strong storms per season develop in the storm track region over the

174   Pacific (Atlantic) Ocean (Table 1).

175   The statistics of a large number of the cyclone trajectories describe the main properties of

176   the Northern Hemisphere storm tracks, including the track density, genesis density, lysis density,

177   and mean storm track intensity. Following Hoskins and Hodges (2002), the track density statistic

178   is calculated by using a single point from each track nearest to each estimation point for each PV

179   cyclone trajectory; the genesis density statistic uses the first detected positions of the cyclones;

180   likewise, the lysis density statistic uses the last detected positions of the cyclones, and the spheri-

181   cal kernel density estimator method (Hodges 1996; Hodges 2008). The genesis and lysis densi-

182   ties are computed as probability density functions (pdf) and scaled to number densities (per unit

183   area per month) by multiplying by the number of points and scaling to a unit area equivalent to a

184   $5°$ spherical cap ($\sim 10^6$ km^2); in the case of the track density, the raw statistic is not a pdf but is

185   scaled to number density by multiplying by the number of tracks and scaled to a unit area equiv-

186   alent to a $5°$ spherical cap. The mean intensity statistic is calculated using a kernel regression es-

187   timator (Hodges 1996) applied to the PV intensity for all points along the cyclone trajectories.

188   For both the density and regression estimators adaptive smoothing is used (Hodges 1996).

189   Sensitivity tests were carried out to assess the robustness of the results in relation to (a)

190   the isentropic level of the analysis on which to describe the storm tracks and (b) the PV intensity

191   threshold, used for the initial identification, above which to consider a cyclone (not shown). An

192   analysis of storm tracks on different isentropic surfaces (not shown) resulted in the choice of the

193   320 K isentrope as it is a good intermediate level on which the storm track features are best rep-

194   resented. The structures and relative intensities of the storm tracks exhibit a lack of sensitivity to

195   the PV intensity threshold (not shown).

196   In general, the storm tracks and the diabatic heating in the corresponding regions act

197   symbiotically in that the presence of the heating helps to maintain the baroclinicity needed for

198   cyclone activity, which in turn influences the 3-dimensional diabatic heating distribution

199   (Hoskins and Valdes 1990). With this co-dependence between the storm tracks and diabatic heat-

200   ing in mind, our study explores the direct relationship between the heating and the storm tracks

201   that influence North America's weather. The diabatic heating is computed diagnostically at each

202   level between 900 and 100-hPa as the residual in the thermodynamic equation (e.g., Hoskins et

203   al. 1989; Barlow et al. 1998; Holton 2004):

204
$$\frac{\dot{Q}(x,y,p,t)}{c_p} = \frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T + \omega \left(\frac{\partial T}{\partial p} - \frac{RT}{c_p p} \right), \qquad (1)$$

205   where $\dot{Q}/c_p$ is the residual heating (K day^{-1}), T the temperature, $\mathbf{v}$ the horizontal wind vector, ω

206   the vertical wind in pressure coordinates, R the gas constant for dry air, c_p the specific heat for

207   dry air at constant pressure, and p the pressure level. The residual is then vertically averaged to

208   yield daily diabatic heating estimates of the free atmosphere.

209   To establish the relationship between the storm tracks and diabatic heating, near-surface

210   winds, and precipitation, we follow a similar approach to that discussed in Hawcroft et al. (2012)

211   and related literature. Each variable is considered to be associated with a cyclone if it is found

212   within a particular circular area around the cyclone center. Precipitation from both reanalysis and

213   observations is considered to be associated with a storm if it is found within a 12° circular area

214   around each storm center, as this is a typical storm precipitation footprint size in the Northern

215   Hemisphere winter (Hawcroft et al. 2012). The reanalysis precipitation is associated with storm

216   centers identified at corresponding 6-hourly time steps, while the GPCP observations are associ-

217   ated with storm center positions at 12:00 UTC each day. Other variables have been reported to

218   be greatly affected within the core of a cyclone represented by a 5° cyclone radius (Hawcroft et

219   al. 2012, 2015), and this is the choice we consider for diabatic heating and near-surface winds

220   which are associated with the storm centers every 6 hours. The storm-related heating, winds, and

221   precipitation fields in the figures are masked out at grid points where the average number of

222   storms is below some very low number (in this case 0.5 storms per unit area per month) in order

223   to highlight the mid-latitude main activity storm track regions.

224

225   c.   *Storm Loss Metrics*

226   To examine the relation between strong-storm tracks and high wind speeds that could

227   lead to potential damage at the surface, we employ a metric defined by Klawa and Ulbrich

228   (2003). The metric is represented by a loss index that highlights areas where strong storms are

229   likely to produce considerable damage by way of winds that exceed the local 98th percentile. Fol-

230   lowing Klawa and Ulbrich (2003),

$$231 \qquad \text{Loss Index} = \sum_{seasons} N_{pop} \left(\frac{v}{v_{98}} - 1 \right)^3 \quad \text{for} \quad v \geq v_{98}, \qquad (2)$$

232   where N_{pop} is the local population number, v the local wind speed related to the storm

233   tracks, and v_{98} the local wind speed at the 98th quantile for 1980-2010. Use of this metric has led

234   to the successful reproduction of storm loss in Germany at the end of the 20th century, in turn

235   leading to a storm loss risk assessment for the nation in the 21st century (Klawa and Ulbrich

236   2003; Donat et al. 2011).

237   Precipitation and storm severity are inherently linked in part by condensational heating

238   and the enhancement of moisture flux convergence (Trenberth et al. 2003). To our knowledge

239 and unlike for winds, a general relationship between storm track precipitation and damaging ef-

240 fects has not yet been established. We adopt a simple approach in which we assume that areas

241 that are most likely to experience loss are those where the storm track precipitation exceeds the

242 local 98th percentile.

243

244 3. DYNAMICS OF STORM TRACKS

245 *a. Environment*

246 The Eady Growth Rate, used in this study, combines information of the static stability

247 and the wind vertical shear for the layer 850-700 hPa, and it is frequently used as a measure of

248 baroclinic instability (Lindzen and Farrell 1980). Following Hoskins and Valdes (1990), Fig. 3

249 shows that (1) regions of large baroclinic instability are found over the western Pacific and At-

250 lantic Oceans (Fig. 3a), and (2) the largest region of low-level baroclinic instability lies poleward

251 of the 200-hPa jet stream (Fig. 3b). Note that a region of high instability in the western Pacific is

252 zonal in orientation and parallels the strong 200-hPa jet. In the western Atlantic, the region of

253 lower troposphere instability also parallels the local upper-level jet maximum with a SW-NE ori-

254 entation that follows the eastern North American coastline. A secondary region of baroclinic in-

255 stability is found in the southeast of the Mediterranean Sea and is also poleward of the corre-

256 sponding local upper-level jet maximum.

257

258   *b.   Physical Properties of the Storm Tracks*

259   The characteristics of the 1980-2010 strong winter storm tracks (as stated, those with

260     $PV_{max} \geq 4.8$ PVU) are depicted in Fig. 4. In Fig. 4a, the mid-latitude trajectories of individual

261   strong storms converge into quasi-zonal bands of high cyclonic activity that form the strong-

262   storm tracks. The number of individual strong storms per unit area, or strong-storm track density

263   (Fig. 4b), is largest over the Pacific, North American-Atlantic (NAA), and Mediterranean re-

264   gions. As expected, and in agreement with Wallace et al. (1988) and Hoskins and Valdes (1990),

265   the three regions of strong-storm tracks are concentrated poleward of the upper-level jets where

266   there is amplified cyclonic shear and enhanced downstream development of baroclinic disturb-

267   ances (Figs. 3b, 4b). Fig. 4b also highlights that the strong-storm track density (shades) resem-

268   bles the track density for all winter storms (as stated, those with $PV_{max} \geq 1$ PVU, the threshold

269   for all-storm tracks, contours), the latter of which is consistent with those presented in Hoskins

270   and Hodges (2002) and other studies. This is particularly evident over the North Atlantic where

271   the NAA storm tracks for both strong storms and all storms extend northeastward from central

272   North America into the higher latitudes near Iceland. The mean intensity statistic denotes the av-

273   erage strength of the strong-storm tracks identified in DJF (Fig. 4c). The strong-storm tracks are

274   most intense where the corresponding track densies are highest (i.e., in the eastern North Pa-

275   cific and western North Atlantic Oceans, and the Mediterranean Sea). The Pacific strong-storm

276   track intensity (shades) shows an eastward shift relative to the corresponding all-storm track

277   (contours). The substantial increase in the strength of strong Pacific storms towards the eastern

278   ocean is indicative of their potential destructive power as they move eastward and hit the North

279   American west coast. Unlike the Pacific track, the NAA strong-storm track retains its high inten-

280   sity across its respective ocean basin. This suggests that the colocation of the low-level baro-

281   clinic zone with the highly active NAA strong-storm track helps to invigorate intense storms in

282   the western Atlantic; in turn, the storms act to reinforce the intensity of the storm track as they

283   propagate across the ocean.

284   Figs. 4d-e illustrate the general temporal evolution of strong storms (shades) that follow

285   the storm tracks. The genesis density statistic in Fig. 4d highlights regions of cyclogenesis, i.e.,

286   the location of the strong storms' initial development. Regions of strong-storm decay are repre-

287   sented by the lysis density statistic (Fig. 4e). Corresponding characteristics of the all-storm tracks

288   are also shown by contours in Figs. 4d-e to display the similarity in behavior between the all-

289   storm and strong-storm tracks.

290   Strong storms that can affect North American weather tend to develop in small groups

291   near low-level baroclinic zones westward of where the storm tracks peak in intensity (Figs. 3a,

292   4d). The storms propagate eastward and become strongest over the eastern North Pacific and

293   western North Atlantic Oceans (Fig. 4c). As they continue to move eastward the strong storms

294   tend to decay (Fig. 4e), in part as they encounter high orography and become disorganized and

295   either dissipate or reorganize leeward of the orography and reinvigorate (Fig. 4d-e). Fig. 4d also

296   shows and supports that strong storms, e.g., intense winter nor'easter storms, which in part are

297   influenced by heat fluxes over the Gulf Stream, tend to develop over the western North Atlantic

298   Ocean near the northeastern United States (Kuo et al. 1991; Davis and Dolan 1993; Yao et al.

299   2008).

300   In the analysis of strong-storm tracks that influence North American weather, it is desira-

301   ble to take into account the corresponding patterns of diabatic heating for the atmospheric col-

302   umn. Figs. 5a-c present the diabatic heating climatology, the heating during all storm activity,

303   and the heating during strong storm activity, respectively. The climatology shows positive heat-

304   ing rates in the western North Pacific and western North Atlantic Oceans (Fig. 5a), and this pat-

305   tern resembles the low-level baroclinic instability (Figs. 3a). The distribution of positive heating

306   rates in the Northern Hemisphere winter is influenced by the distribution of the warm Kuroshio

307   and Gulf Stream currents in the western North Pacific and North Atlantic Oceans, respectively,

308   and by the zonal asymmetry of the land-ocean distribution (Brown 1964; Geller and Avery 1978;

309   Wei et al. 1983). In contrast to the climatology, the heating during all storm activity increases in

310   strength and spreads across the North Pacific and North Atlantic Oceans in the mid-latitudes

311   (Fig. 5b). The heating is even more intense during strong storm activity (Fig. 5c). In the North

312   Pacific, the heating further intensifies in the east where the Pacific strong-storm track is most in-

313   tense, and it remains strong as it spreads up and down the west coast of North America. In rela-

314   tion to the NAA strong-storm track, the heating is most intense over the western North Atlantic

315   and remains strong across the ocean where the storm track retains its high intensity.

316   Fig. 5d presents the ratio of the positive heating rates related to strong-storm activity to

317   the positive heating rates related to all-storm activity. This comparison between the strong-storm

318   and all-storm heating reveals that the heating related to the strong-storm tracks is at least 25%

319   more intense than the heating related to the all-storm tracks over the Pacific and Atlantic Oceans

320   where the storm tracks are strongest. Moreover, in the lower mid-latitudes, the strong-storm heat-

321   ing is up to 3 times more intense than the all-storm heating.

322   Deep convection associated with strong-storm activity is obtained directly as a diagnostic

323   from the CFSR database (Fig. 6). High positive heating rates associated with deep convection are

324   found in each of the strong-storm track regions and are highest where the storm tracks are most

325   intense (see Figs. 4b-c). Furthermore, the heating from deep convection largely resembles the di-

326   abatic heating distribution in the strong-storm track regions (Figs. 5c), suggesting that deep con-

327   vective processes dominate the strong-storm tracks in the free atmosphere.

328   The strong-storm diabatic heating in the western North Atlantic corresponds with the

329   higher track density and is more intense than the heating in the North Pacific (Figs. 4b, 5c). Simi-

330   lar relationships are found in the deep convection associated with strong-storm tracks (Fig. 6).

331   Along with the local SW-NE oriented low-level baroclinic zone and upper-level jet near the east

332   coast of North America (Fig. 3), the stronger heating in the Atlantic promotes greater instability

333   and increased cyclonic activity (Fig. 4b), supporting the findings of Brayshaw et al. (2009).

334

335   4.   RELATION OF STORM TRACKS TO SURFACE WEATHER

336   The near-surface wind distribution can change dramatically during the evolution of in-

337   tense extratropical cyclones, and this is cause for concern for two reasons. First, in populated

338   areas there is great potential for the wind to inflict serious damage and put lives in jeopardy, and

339   second, over open waters strong near-surface winds have great impacts where maritime transpor-

340   tation, fishing vessels, and manned offshore oil and gas production units are most vulnerable

341   (Bell et al. 2017). Strong storms can also change the winter precipitation distribution by generat-

342   ing excessive amounts in a relatively short amount of time (days to a week). Lasting and possibly

16

343   devastating effects like major flooding and wind damage may result leading to states of emer-

344   gency, especially when the cumulative impacts and insurance losses from several storms occur-

345   ring in rapid succession are considered (Mailier et al. 2006). The patterns of intense near-surface

346   winds and heavy precipitation rates in strong-storm tracks over North America are explored next.

347

348   *a.   Near-surface Winds*

349   The relation between the storm tracks and near surface winds will be discussed next with

350   the support of Figs. 7 and 8. In the absence of storm activity (Fig. 7a), near-surface winds

351   achieve magnitudes of about 4-6 m s^{-1} and resemble the pattern of the upper-level jet presented

352   in Fig. 3b with primarily eastward and northeastward directions in the North Pacific and North

353   Atlantic Oceans, respectively. Fig. 7b shows that for both the Pacific and NAA all-storm tracks

354   (i.e., storms with $PV_{max} \geq 1$ PVU), the near-surface winds intensify where the storm tracks are

355   strongest and shift eastward in the eastern ocean basins. The winds associated with the strong-

356   storm tracks (i.e., $PV_{max} \geq 4.8$ PVU) presented in Fig. 7c show further intensification and a

357   stronger eastward shift over the oceans.

358   The increases in wind speed related to the all-storm tracks are better seen in Figs. 8a and

359   8b that depict the difference and ratio, respectively, between the all-storm related wind speeds

360   and the no-storm wind speeds. Likewise, Figs. 8c and 8d depict the wind speed difference and

361   ratio between strong-storm and no-storm events. In the North Pacific Ocean, wind speeds in-

362   crease in the eastern basin where the all-storm track is strongest (Fig. 8a), particularly in the

363   lower and higher mid-latitudes where they are over 5 times more intense (Fig. 8b). The winds

364   over the ocean further intensify during strong storm activity (Figs. 8c-d), helping to drive strong

365   storms eastward to the North American coast. Moving to the Atlantic sector, wind speeds are

366   found to intensify across the North Atlantic but particularly in the west and lower mid-latitudes

367   with a secondary maximum towards the northeastern sector (Figs. 8a-b; also seen in Fig 7c).

368   During strong storm activity, wind speeds are further enhanced, specifically in the west just off-

369   shore of North America (Figs. 8c-d). Over land, near-surface wind speeds related to the all-storm

370   tracks increase east of the Rocky Mountains (Figs. 8a-b). Greater intensification in the wind

371   speeds is evident during strong-storm events (Figs. 8c-d), specifically in the eastern US where

372   the corresponding strong-storm track strengthens (see Fig. 4c).

373   Overall the strong-storm tracks leave greater imprints in the near-surface wind field in the

374   North Atlantic than in the North Pacific (Fig. 7c), most notably just offshore of North America

375   where maritime shipping and oil platforms are exposed. Increases in wind speeds near the coasts

376   are also more pronounced in the western North Atlantic (Figs. 8c-d), consistent with the distribu-

377   tions of diabatic heating and deep convection that indicate greater baroclinic instability in the re-

378   gion (see Figs. 5c, 6). This would suggest that maritime operations in the western North Atlantic

379   are more at risk to damage by way of near-surface winds associated with the strong-storm tracks.

380   The potential damage associated with extratropical strong-storm tracks over land in North

381   America is assessed taking into account very high near-surface storm wind speeds, i.e., those that

382   exceed the local 98th percentile (Fig. 9). Areas east of high orography experience the highest per-

383   cent of strong-storm days with near-surface wind speeds above the 98th percentile (Fig. 9a). As

384   seen in Fig. 9b, these same areas also experience the most intense wind speeds related to the

385   strong-storm tracks.

386   Intense near-surface winds do not necessarily imply damage, unless they occur over pop-

387 ulated areas. Fig. 9c presents the 2010 population number obtained from the LandScan Global

388 Population Project following the methodology in Dobson et al. (2000), which is used for the cal-

389 culation of the storm loss index presented in section 2c. The storm loss estimate (Fig. 9d) high-

390 lights the regions that are most vulnerable to damages from very high storm winds. Within North

391 America, these areas are in the eastern US spanning from the Midwest to the east coast states, as

392 well as along the southwestern US coast. A comparison with Fig. 9a reveals that storm wind loss

393 in these areas is associated with up to 16% of strong storms in winter.

394

395 *b.   Precipitation*

396   Figure 10 presents the relation between the storm tracks and surface precipitation. Figs.

397 10a and 10b show the precipitation distributions related to all-storm tracks and to strong-storm

398 tracks, respectively. Consistent with the findings in Hawcroft et al. (2012), the all-storm and

399 strong-storm precipitation maxima are found in the North Pacific and North Atlantic Oceans as

400 well as along the west coast of North America. Secondary precipitation maxima are found in the

401 southeastern United States. Overall the storm tracks leave greater imprints in the precipitation in

402 the North Atlantic than in the North Pacific, possibly associated with the warmer waters that fa-

403 vor increased baroclinic instability and deep convection (see Figs. 3a, 6).

404   The difference and ratio between the strong-storm and all-storm track precipitation (Figs.

405 10c and 10d respectively) indicate the noticeable increases in precipitation that result from the

406 fewer but stronger storms. These increases are evident over the oceans where the storm tracks are

407 most intense. The percent contribution of strong-storm precipitation to the all-storm precipitation

408   (Fig. 10e) shows that strong storms represent about 16% of all storms, yet they contribute 30-

409   50% of the precipitation associated with the Pacific and NAA storm tracks (discussed further be-

410   low).

411   As expected, precipitation associated with strong-storm activity in the eastern Pacific

412   Ocean is more intense than that associated with all-storm activity (Figs. 10a-c). Towards the US

413   west coast, precipitation increases during strong-storm activity (Figs. 10c-d) and contributes to

414   almost half of the all-storm precipitation in the region (Fig. 10e), exposing local fishing and

415   other maritime operations to potential damage. Further increases are found as North America's

416   land contrasts and orographic effects come into play: in the western US, increases of 50% are

417   found during strong storm activity (Figs. 10d-e). This supports the notion that cyclones aloft lead

418   to large accumulations of precipitation upstream of great mountain ranges and other high orogra-

419   phy (Garreaud 2007). Farther east, strong-storm tracks are also associated with more intense pre-

420   cipitation rates (Fig. 10c-d), contributing to over 30% of the all-storm precipitation (Fig. 10e).

421   Orography in the northeastern United States can further boost the precipitation from strong

422   storms. Similar results are found in the western North Atlantic where the enhanced strong-storm

423   precipitation contributes to 30% of the all-storm precipitation (Figs. 10c-e).

424   The susceptibility to damage from heavy precipitation, i.e., precipitation rates exceeding

425   the local 98th percentile, during strong storm activity is investigated for North America (Fig. 11).

426   The percent of strong-storm days with precipitation rates exceeding the local 98th percentile is

427   presented in Fig. 11a. Distributions of heavy precipitation related to the all-storm and strong-

428   storm tracks are shown in Figs. 11b and 11c, respectively. The heaviest precipitation related to

429   the all-storm and strong-storm tracks is found along the west coast and in the southeastern US

430 (Figs. 11b-c). The strong-storm tracks in particular likely play key roles in shaping the precipita-

431 tion distribution in the southeastern US as the region experiences a relatively high percentage of

432 strong-storm days with precipitation exceeding the 98th percentile (Fig. 11a). Along the west

433 coast where there are fewer strong-storm days with heavy precipitation, it is likely that in addi-

434 tion to the strong-storm tracks, other factors such as topography and land-ocean contrasts may

435 influence the distribution of heavy precipitation. In the central US, the high percentage of strong-

436 storm days with heavy precipitation corresponds to low strong-storm precipitation rates (Figs.

437 11a,c). This indicates that in winter the region is relatively dry during strong-storm activity and

438 is therefore less likely to experience loss associated with precipitation exceeding the 98th percen-

439 tile.

440   Fig. 11d highlights the differences between the heavy strong-storm and all-storm track

441 precipitation, and Fig. 11e presents the percent contribution of strong storms to all-storm precipi-

442 tation that exceeds the 98th percentile. Substantial increases in precipitation rates during strong

443 storm activity are found in the southeastern US and near the US east coast (Fig. 11d). Areas with

444 the largest increases in heavy precipitation correspond to regions where strong storms contribute

445 well over 30% of the all-storm precipitation (Fig. 11e), indicating their vulnerability to damage

446 related to heavy strong-storm precipitation. The southeastern US is particularly vulnerable as

447 precipitation is greatly enhanced during strong storm activity and contributes almost 50% of

448 heavy all-storm precipitation in the region.

449

450 *c.   Reanalysis vs. observed precipitation related to the Storm Tracks*

451   The precipitation blending algorithm in CFSR combines pentad Climate Prediction Cen-

452   ter (CPC) Merged Analysis of Precipitation (CMAP) and daily gauge precipitation analyses of

453   varying spatial resolutions with background 6-hourly precipitation from the Global Data Assimi-

454   lation System, GDAS (Saha et al. 2010). The blending algorithm in CFSR is latitude dependent:

455   in the tropics it tends to the CMAP analysis, in the mid-latitudes to a gauge analysis, and in the

456   high latitudes to the model precipitation. Therefore, despite CFSR including precipitation in its

457   assimilation cycle, deviations from observations may occur. During 1999-2010, daily GPCP pre-

458   cipitation rates are considerably less intense than the daily reanalysis precipitation rates (not

459   shown, but almost identical to the 1980-2010 reanalysis precipitation rates), particularly north of

460   60°N along the southern coastlines of Alaska and Greenland. As stated, this and other differences

461   in winter precipitation between CFSR and GPCP may be due to multiple reasons, including the

462   precipitation blending algorithm in CFSR but also inadequate satellite-driven estimations of pre-

463   cipitation at high latitudes included in the daily GPCP dataset (Bolvin et al. 2009).

464   We examine whether the relation of the strong-storm tracks with the daily reanalysis pre-

465   cipitation is maintained over North America when using precipitation derived from observations,

466   that is, the daily precipitation from GPCP (section 2). To this end, and despite that GPCP became

467   available in 1996, the period 1999-2010 is examined to avoid any eventual spurious effects due

468   to the 1998 discontinuity found in CFSR. The 1999-2010 daily precipitation distributions associ-

469   ated with the all-storm and strong-storm tracks for GPCP are shown in Figs. 12a and 12b, respec-

470   tively. Comparison with the reanalysis precipitation (Figs. 10a-b) indicates that they share simi-

471   lar spatial distributions with local maxima over the eastern North Pacific Ocean, the western

472   North Atlantic Ocean, the west coast of North America, and the southeastern United States. Nev-

473   ertheless, the GPCP precipitation does exhibit weaker intensities, particularly in the Pacific and

474 NAA storm track regions over the oceans. It is likely that the discrepancy in magnitude results

475 from uncertainties in the oceanic observations of precipitation described in Adler et al. (2012).

476   Similar inferences can be noted in the difference (Fig. 12c) and ratio (Fig. 12d) of the ob-

477 served precipitation related to the all-storm and strong-storm tracks. Differences in the reanalysis

478 and observed precipitation metrics are noted particularly in the western North Atlantic Ocean

479 where the observed precipitation related to the strong-storm tracks is shown to decrease (Fig.

480 12c). As already stated, the uncertainties in oceanic observations may play a role in this discrep-

481 ancy. Over land, the observed precipitation differences and ratios in Figs. 12c and 12d show in-

482 creases along the US west coast and in the southeastern US, consistent with the reanalysis (see

483 Figs. 10c-d). The contribution of strong storms to the observed all-storm precipitation is pre-

484 sented in Fig. 12e. As depicted in the reanalysis (Fig. 10e), observations show that strong storms

485 contribute over 30% of the all-storm precipitation over land and the oceans.

486   We also analyze the relation of the storm tracks with precipitation from GPCP that ex-

487 ceeds the local 98th percentile in North America (Fig. 13). Comparison of the reanalysis (Figs.

488 11b-e) and observational metrics reveal similarities despite the weaker GPCP intensities. The in-

489 tense precipitation observed over the continent (Figs. 13a-b) corresponds qualitatively well with

490 the reanalysis, in particular in the eastern US and along the North American west coast where the

491 precipitation is further enhanced during strong storm activity (Fig. 13c). According to Fig. 13d,

492 strong storms contribute over 30% of the all-storm precipitation that exceeds the 98th percentile

493 in regions where large increases are observed. The results indicate that the eastern US and the

494 west coast of North America are most prone to damage from heavy strong-storm precipitation,

495   consistent with the findings using CFSR (section 4b). In general, we find that the reanalysis pre-

496   cipitation distributions related to the all-storm and strong-storm tracks are consistent with obser-

497   vations.

498

499   *d.   The 1998 CFSR data discontinuity and the Storm Tracks*

500   It was earlier stated that the reanalysis data show a discontinuity in the wind and precipi-

501   tation fields in October 1998 thought to be due to the ingestion of data from ATOVS at the time.

502   For instance, after 1998 there is a marked decrease in the intensity of low-level winds in the trop-

503   ics and an increase in the global average precipitation (Chelliah et al. 2011; Wang et al. 2011).

504   We investigate what impact, if any, this jump has on the results. To this end, the subset periods

505   of 1980-1998 (hereafter, the early period) and 1999-2010 (hereafter, the later period) are ana-

506   lyzed. Table 1 displays relevant strong-storm statistics for the early and later periods to assess

507   any change in the strong winter storm tracks that could impact the North American climate. The

508   statistics are normalized to units per season and include the number of strong storms identified,

509   the mean intensity of the strong storms, and the average maximum intensity reached by the

510   strong storms during each period. Furthermore, each decade between 1980 and 2010 is examined

511   to explore the possibility of a trend in the storm tracks regardless of the discontinuity.

512   The more important feature noted in Table 1a is that no noticeable variations are found in

513   the statistical means between the early and later periods and among the decades within 1980-

514   2010. This indicates that the CFSR discontinuity does not significantly influence NH storm track

515   behavior. Further, the effect of the discontinuity on the Pacific and NAA strong-storm tracks sep-

516   arately is investigated (Tables 1b-c), and it is found again that the behavior of each of the storm

517   tracks is unaffected.

518   A related evaluation was performed for the relation between the strong-storm tracks and

519   the near-surface wind and precipitation distributions (not shown). Again, it was found that the

520   1998 CFSR discontinuity has little or no influence on the results corresponding to North Ameri-

521   can high impact weather. The wind speed associated with strong-storm tracks in each of these

522   periods resembles that for the entire period and the same is true for the strong-storm precipita-

523   tion. In summary, it is found that the CFSR discontinuity does not affect any of the features dis-

524   cussed in this article.

525

526   5.   CONCLUDING REMARKS

527   The behavior of strong winter storm tracks and their imprint on storm track-related

528   weather in North America are discussed using 31 years of data from the Climate Forecast System

529   Reanalysis and 12 years of precipitation data from the Global Precipitation Climatology Project.

530   It is found that a data discontinuity in October 1998 in CFSR does not affect the behavior of the

531   Northern Hemisphere storm tracks, nor does it influence their relation with North American win-

532   ter weather. Storms are defined as maxima in potential vorticity and objectively tracked through

533   their lifecycles following a Lagrangian approach. Two types of storm tracks are discussed: the

534   first one, "all-storm tracks", includes all extratropical cyclones whose maximum PV intensities

535   exceed a low threshold of 1 PVU; the second type, "strong-storm tracks", only includes storms

536   that achieve a maximum potential vorticity of at least 4.8 PVU, which is the value exceeding the

537 mean intensity of storms comprising the Pacific and NAA storm tracks by one standard devia-

538 tion. These more intense extratropical cyclones make up about 16% of all winter storms. Both

539 all-storm tracks and strong-storm tracks are found to correspond well with those described in

540 previous studies: over the North Pacific Ocean and over the North Atlantic Ocean (as well as a

541 weaker one over the Mediterranean Sea). In addition to detecting larger structures like the mean

542 intensity of the storm tracks, and because of the dependence of PV on higher order derivatives,

543 small-scale features of the storm tracks are easily differentiated, i.e., regions of cyclogenesis and

544 cyclolysis. The cyclogenesis pattern shows that strong storms generally develop near low-level

545 baroclinic zones. The cyclolysis pattern reveals that the strong storms tend to dissipate in the

546 eastern North Pacific Ocean, the western North Atlantic Ocean near eastern Canada, and a sec-

547 ondary area over the central United States. The symbiotic relation between storm tracks and dia-

548 batic heating is evidenced in the large increases in diabatic heating associated with deep convec-

549 tive processes. The heating increases occur where the strong-storm tracks are most intense, in

550 particular over the oceans.

551   The analysis of the relation of strong-storm tracks to the near-surface wind distribution

552 indicates that the winds shift eastward during strong storm activity. Furthermore, the wind

553 speeds increase over the oceans where the storm tracks are most intense, i.e., in the eastern North

554 Pacific and western North Atlantic Oceans. Over North America, areas east of the Rockies ex-

555 hibit large increases in wind speed during strong storm activity. It is found that the precipitation

556 associated with strong-storm tracks is most intense where they are strongest. Moreover, the pre-

557 cipitation during strong storm activity is more intense than that during all storm activity, espe-

558 cially in the North Atlantic Ocean where the NAA storm track density is particularly high. While

559   strong-storms make up about 16% of all-storms, they contribute 30-50% of the all-storm precipi-

560   tation over the oceans and over North America. Calculations based on an observed precipitation

561   dataset (GPCP) confirm results based only on CFSR products and thus support the robustness of

562   the findings.

563   The analysis of very high wind speeds and heavy precipitation related to the strong-storm

564   tracks provides an inference of their destructive potential in North America. While the most in-

565   tense strong-storm wind speeds are found in the central United States, areas most likely to expe-

566   rience the greatest storm wind-related loss span from the Midwest to the east coast states as well

567   as along the southwestern US coast. Heavy precipitation is further enhanced during strong storm

568   activity, with the largest increases occurring along the west coast, in the southeastern US, and

569   near the US east coast. In these areas, strong storms contribute over 30% of the all-storm precipi-

570   tation that exceeds the local 98th percentile, indicating their vulnerability to damages from heavy

571   precipitation during strong storm activity.

572   Our findings indicate that strong-storm tracks leave a significant imprint on winter

573   weather in North America, despite making up a small fraction of all storms that develop. This

574   imprint depends not only on dynamical features but also on the density of the population, thus

575   showing the greatest loss in the eastern US and North American coasts. Over the water, it would

576   be expected that oil platforms and maritime shipping and fishing craft are most vulnerable to

577   storm-related damages just offshore of the US coasts.

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744 **Table**

745

(a) **Northern Hemisphere**	*1980-2010*	*1980-1998*	*1999-2010*	*1981-1990*	*1991-2000*	*2001-2010*
Number of Strong Storms (season⁻¹)	26	24	29	26	24	30
Mean Intensity (PVU)	3.8	3.8	3.8	3.8	3.8	3.8
Average Max Intensity (PVU)	5.4	5.4	5.4	5.4	5.4	5.4

746

(b) **Pacific storm track**	*1980-2010*	*1980-1998*	*1999-2010*	*1981-1990*	*1991-2000*	*2001-2010*
Number of Strong Storms (season⁻¹)	9	9	9	10	7	9
Mean Intensity (PVU)	3.8	3.8	3.8	3.8	3.7	3.8
Average Max Intensity (PVU)	5.3	5.3	5.3	5.3	5.3	5.3

747

(c) **North American-Atlantic storm track**	*1980-2010*	*1980-1998*	*1999-2010*	*1981-1990*	*1991-2000*	*2001-2010*
Number of Strong Storms (season⁻¹)	6	5	8	5	6	8
Mean Intensity (PVU)	3.8	3.8	3.8	3.8	3.8	3.9
Average Max Intensity (PVU)	5.3	5.3	5.3	5.3	5.3	5.3

748

749

750 **Table 1**: Statistics for the DJF strong-storm tracks for (a) the entire Northern Hemisphere, (b) the Pacific storm track, and (c) the

751 North American-Atlantic storm track. In (b) and (c), only strong storms that develop within the specified storm track domain are in-

752 cluded. The first column shows the values for the entire 31-year period. The following two columns denote the early and later periods.

753 The last three columns highlight the values for each decade.

Figure Caption List

755 Figure 1: DJF mean zonal state in the Northern Hemisphere for 1980-2010. The mean zonal

756   wind is shaded with 5.0 m s^{-1} intervals. Line contours indicate the vertical distribution of

757   mean zonal isentropic surfaces at a 10 K contour interval. The bold black line highlights

758   the θ=320K surface on which the mid-latitude storm tracks are defined.

759 Figure 2: Histogram of all DJF storms binned by maximum intensity in the Northern Hemisphere

760   for 1980-2010. Maximum intensity bins are shown in the x-direction at an interval of 0.2

761   PVU. Storms included in the all-storm track analysis have maximum intensities of 1 PVU

762   or greater. Strong storms that follow the Pacific (PAC) or North American-Atlantic

763   (NAA) storm tracks have maximum intensities of 4.8 PVU or greater and are highlighted

764   in warm colors. In parentheses in the labels, NH signifies the statistics for the Northern

765   Hemisphere, while ST indicates the statistics for the PAC and NAA storm tracks.

766 Figure 3: (a) 1980-2010 DJF Eady growth rate average for the 850-700 hPa layer. Values ex-

767   ceeding 0.2 day^{-1} are shaded at 0.2 day^{-1} intervals. Masked areas over the continents indi-

768   cate regions where the land extends above the 850-hPa surface. (b) Zonal mean wind at

769   200-hPa. Values exceeding 15 m s^{-1} are shaded.

770 Figure 4: Storm track statistics in the Northern Hemisphere DJF season for 1980-2010. All-storm

771   tracks properties are depicted in contours, while the strong-storm track properties are

772   shaded. (a) Individual trajectories of strong storms; (b) Track density for all-storm tracks

773   (contours at intervals of 3.0 storms per 10e6 km^2 per month) and strong-storm tracks

774   (shaded at intervals of 0.5 storms per 10e6 km^2 per month); (c) Mean intensity of all-

775   storm tracks (contour intervals of 0.4 PVU) and strong-storm tracks (shaded at intervals

776   of 0.2 PVU); (d) cyclogenesis density for all-storm tracks (contours at intervals of 0.4

777   storms per 10e6 km^2 per month) and strong-storm tracks (shaded at intervals of 0.05

778   storms per 10e6 km^2 per month); (e) as (d) but for cyclolysis.

779   Figure 5: DJF 1980-2010 vertically averaged 900-100 hPa diabatic heating: (a) Climatology; (b)

780   during all storm activity; and (c) during strong storm activity. (d) The ratio (%) of the

781   strong-storm diabatic heating to the all-storm diabatic heating. Shaded regions in (d) indi-

782   cate areas where the all-storm and strong-storm heating rates are positive.

783   Figure 6: Mean heating from deep convection during strong storm activity averaged between

784   900-100 hPa in the Northern Hemisphere DJF season for 1980-2010. Contour interval is

785   1.0 K day^{-1}. Regions outside the all-storm track regions are masked out.

786   Figure 7: Mean near-surface wind distributions on the hybrid level 1 in DJF for 1980-2010 (a)

787   during no storm activity, (b) during all storm activity, and (c) during strong storm activ-

788   ity. Shaded intervals are 2.0 m s^{-1}. In (b) and (c), regions outside the all-storm track re-

789   gions are masked out.

790   Figure 8: Wind speed comparisons based on Fig. 7. (a) Difference between all-storm wind speed

791   and no-storm wind speed. (b) Ratio (%) of the all-storm wind speed to the no-storm wind

792   speed. (c) Difference between strong-storm wind speed and no-storm wind speed. (d) Ra-

793   tio (%) of strong-storm wind speed to the no-storm wind speed. In (a) and (c), shaded in-

794   tervals are 1.0 m s^{-1}. In (b) and (d), values exceeding 100% are shaded with intervals of

795   50%. Regions outside the all-storm track regions are masked out.

796   Figure 9: Analysis of intense near-surface wind speeds in DJF for 1980-2010 in North America.

797   (a) Percent of strong-storm days with wind speeds exceeding the local 98th percentile.

798   Shaded intervals are 2%. (b) Mean strong-storm wind speeds exceeding the local 98th

799   percentile. Shaded intervals are 2 m s^{-1}. (c) 2010 population number with an interval of

800   1e4 people. (d) The strong-storm wind speed loss index with an interval of 5e5 and all

801   positive values shaded.

802   Figure 10: Analysis of CFSR precipitation rates (PR) during DJF for 1980-2010. (a) The mean

803   precipitation during all storm activity, and (b) the mean precipitation during strong storm

804   activity. In (a) and (b), shaded intervals are 1.0 mm day^{-1}. (c) The difference between

805   strong-storm precipitation and all-storm precipitation with an interval of 0.5 mm day^{-1}.

806   (d) The ratio (%) of strong-storm precipitation to all-storm precipitation with an interval

807   of 10% and values exceeding 100% are shaded. (e) Percent contribution of strong storms

808   to all-storm precipitation with an interval of 5%. For all panels, areas outside the all-

809   storm track regions are masked out.

810   Figure 11: Analysis of CFSR intense precipitation rates (PR) in DJF for 1980-2010 in North

811   America. (a) The percent of strong-storm days with precipitation exceeding the local 98th

812   percentile. Shaded intervals are 1%. (b) The all-storm precipitation that exceeds the local

813   98th percentile. Shaded intervals are 10 mm day^{-1}. (c) As in (b) but for strong-storm pre-

814   cipitation. (d) The difference between strong-storm precipitation and all-storm precipita-

815   tion. Shaded intervals are 2.0 mm day^{-1}. (e) Percent contribution of strong storms to all-

816   storm precipitation with an interval of 5% and all values exceeding 10% shaded. Masking

817   for all panels indicates areas where storm precipitation falls below the local 98th percen-

818   tile.

819   Figure 12: As in Fig. 10 but for GPCP precipitation for 1999-2010.

820   Figure 13: As in Figs. 11b-e but for GPCP precipitation for 1999-2010.
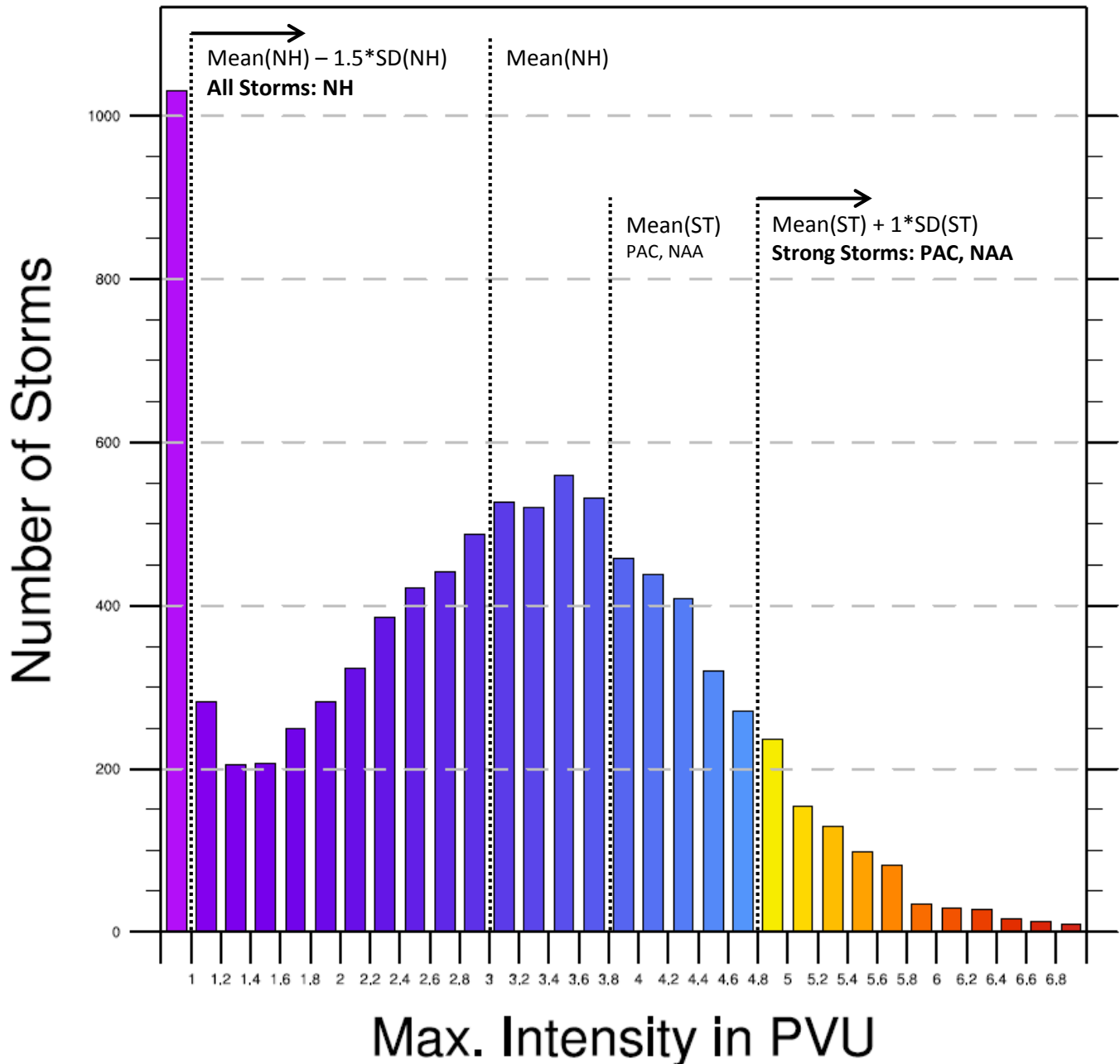
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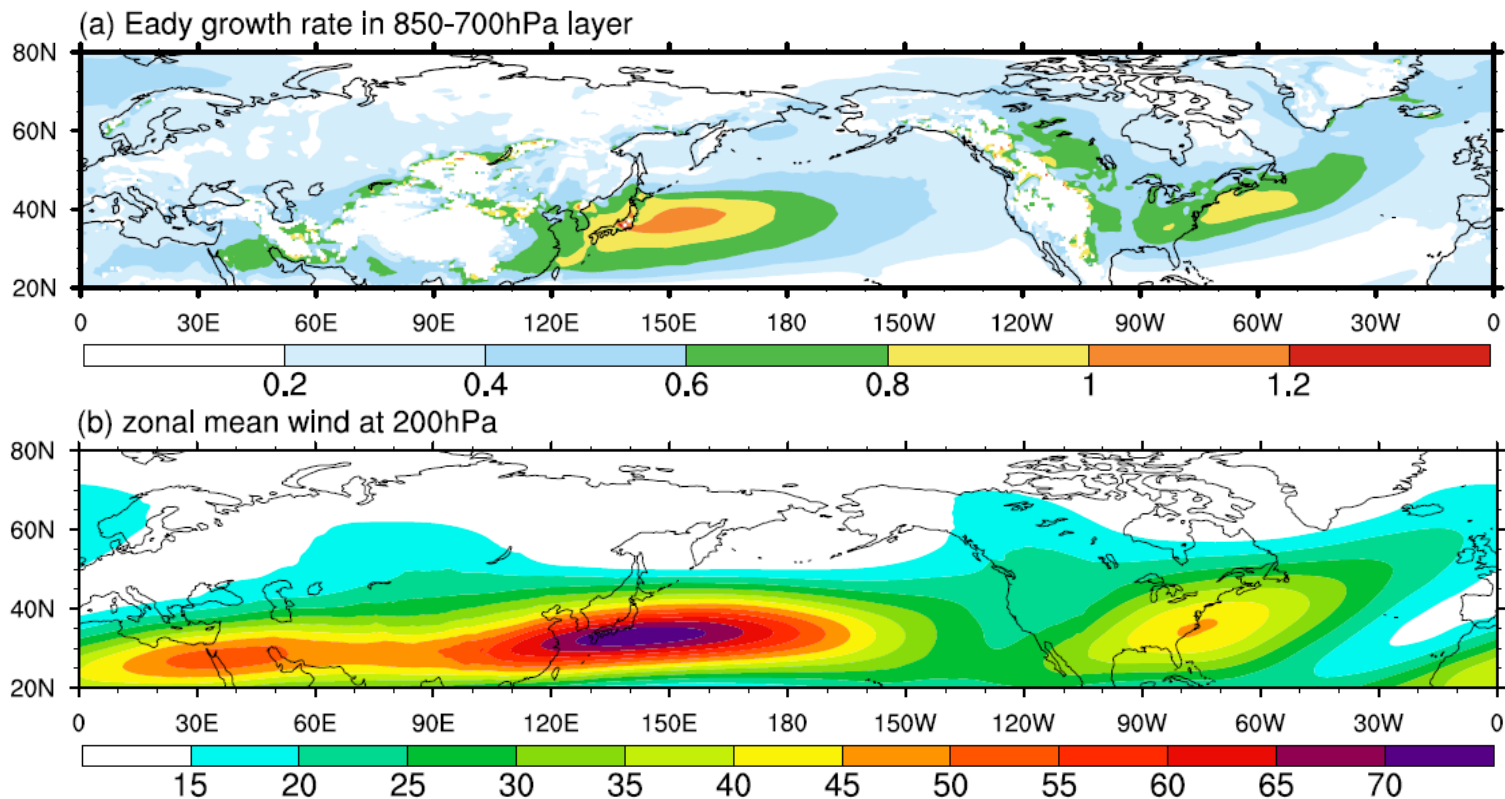
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(a) Eady growth rate in 850-700hPa layer

(b) zonal mean wind at 200hPa

Figure 3: (a) 1980-2010 DJF Eady growth rate average for the 850-700 hPa layer. Values exceeding 0.2 day^{-1} are shaded at 0.2 day^{-1} intervals. Masked areas over the continents indicate regions where the land extends above the 850-hPa surface. (b) Zonal mean wind at 200-hPa. Values exceeding 15 m s^{-1} are shaded.
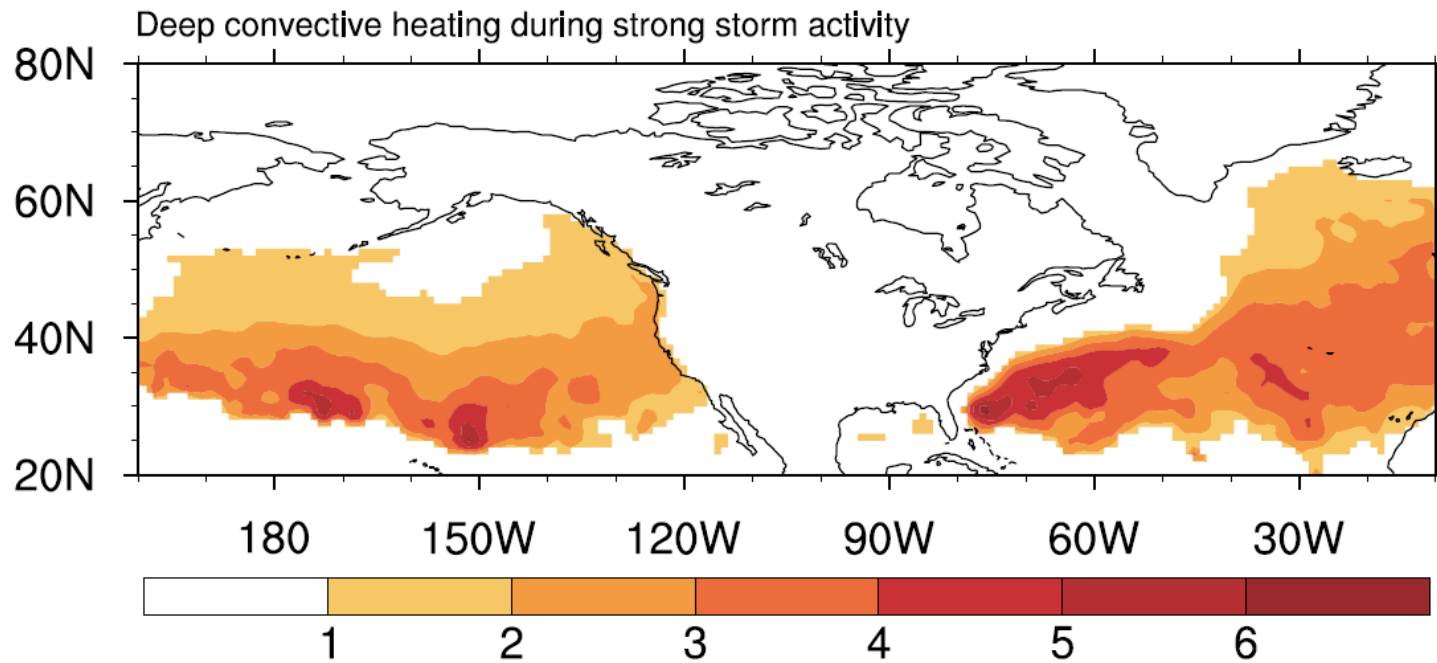
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856

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Deep convective heating during strong storm activity

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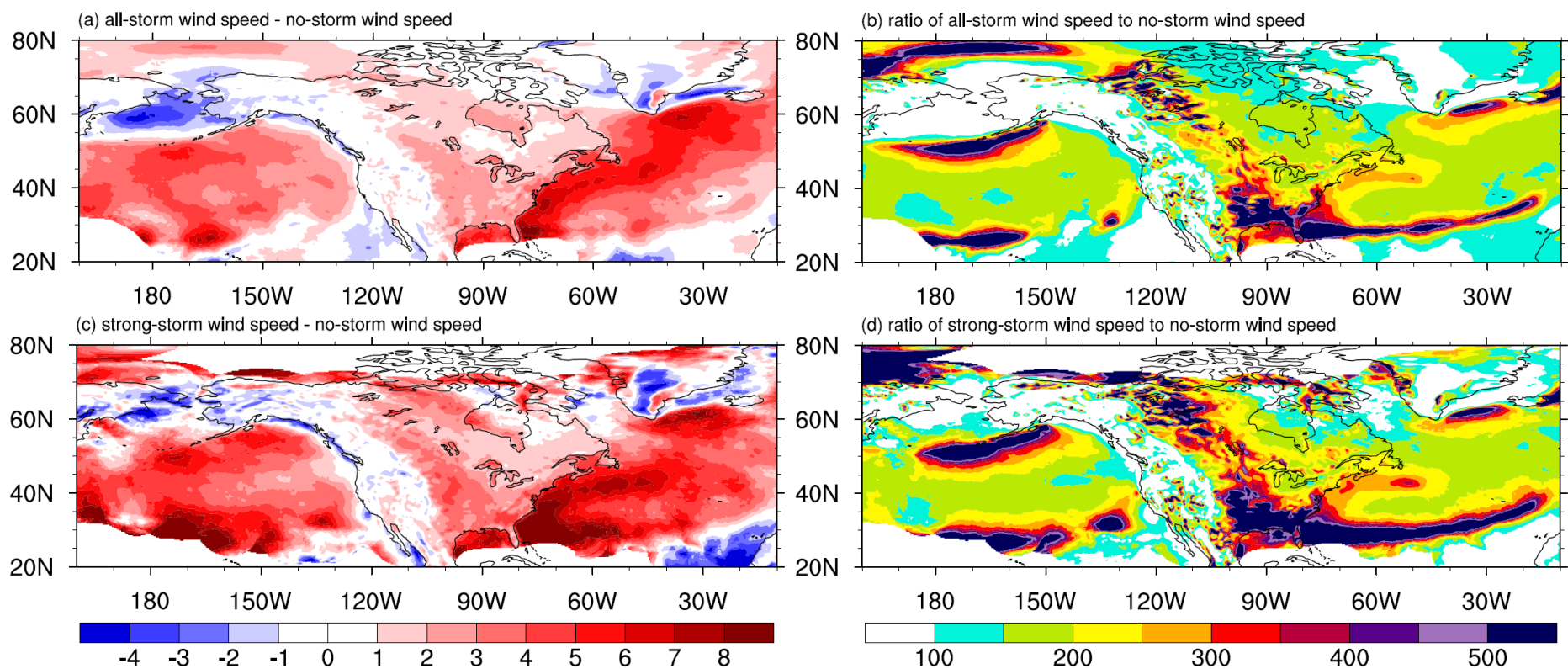
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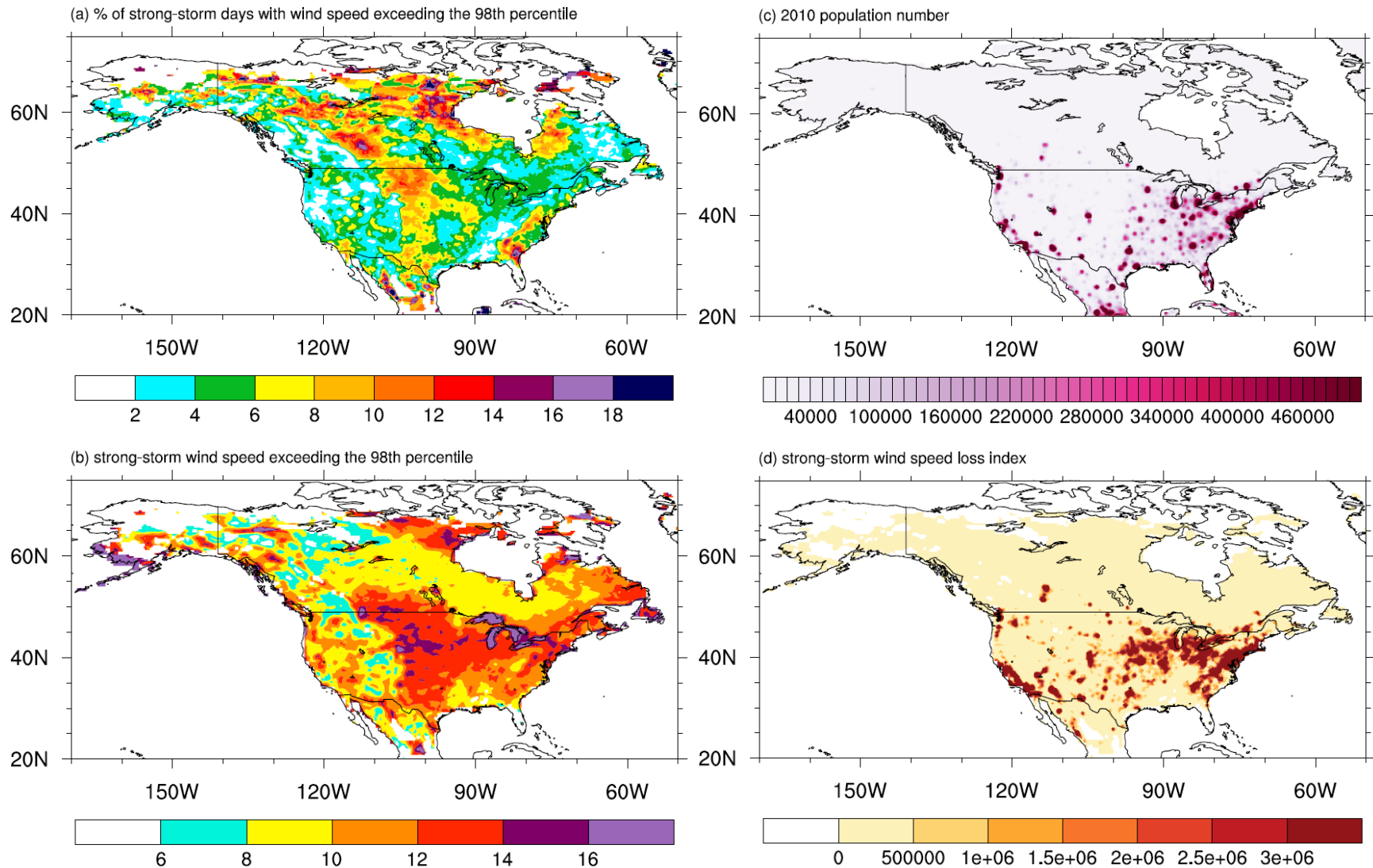
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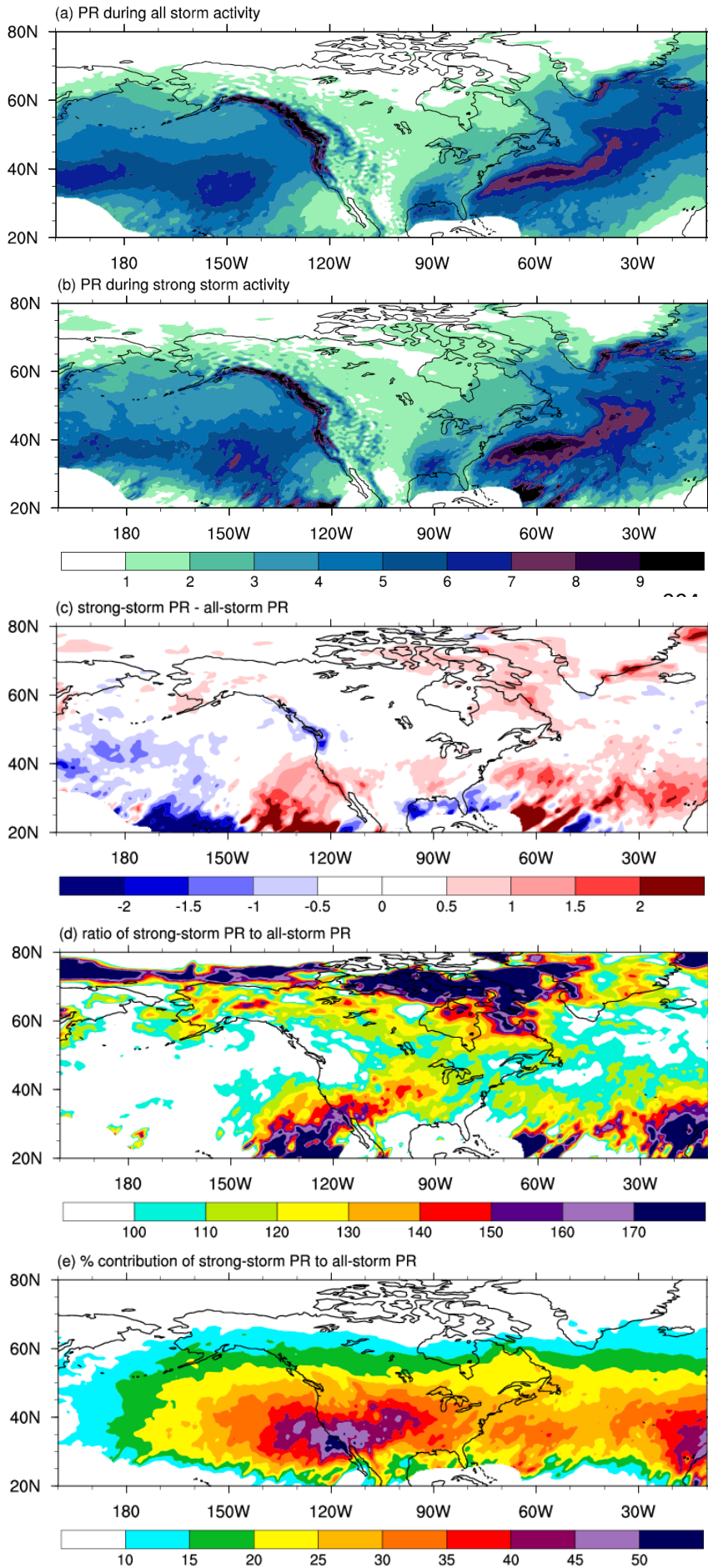
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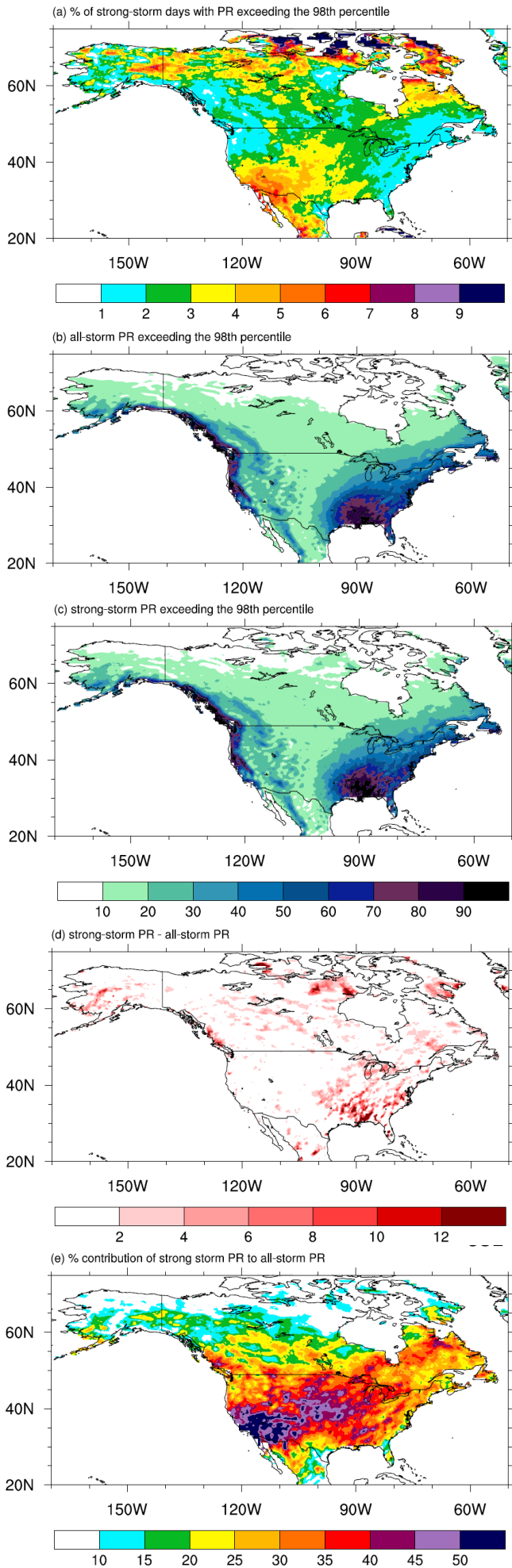
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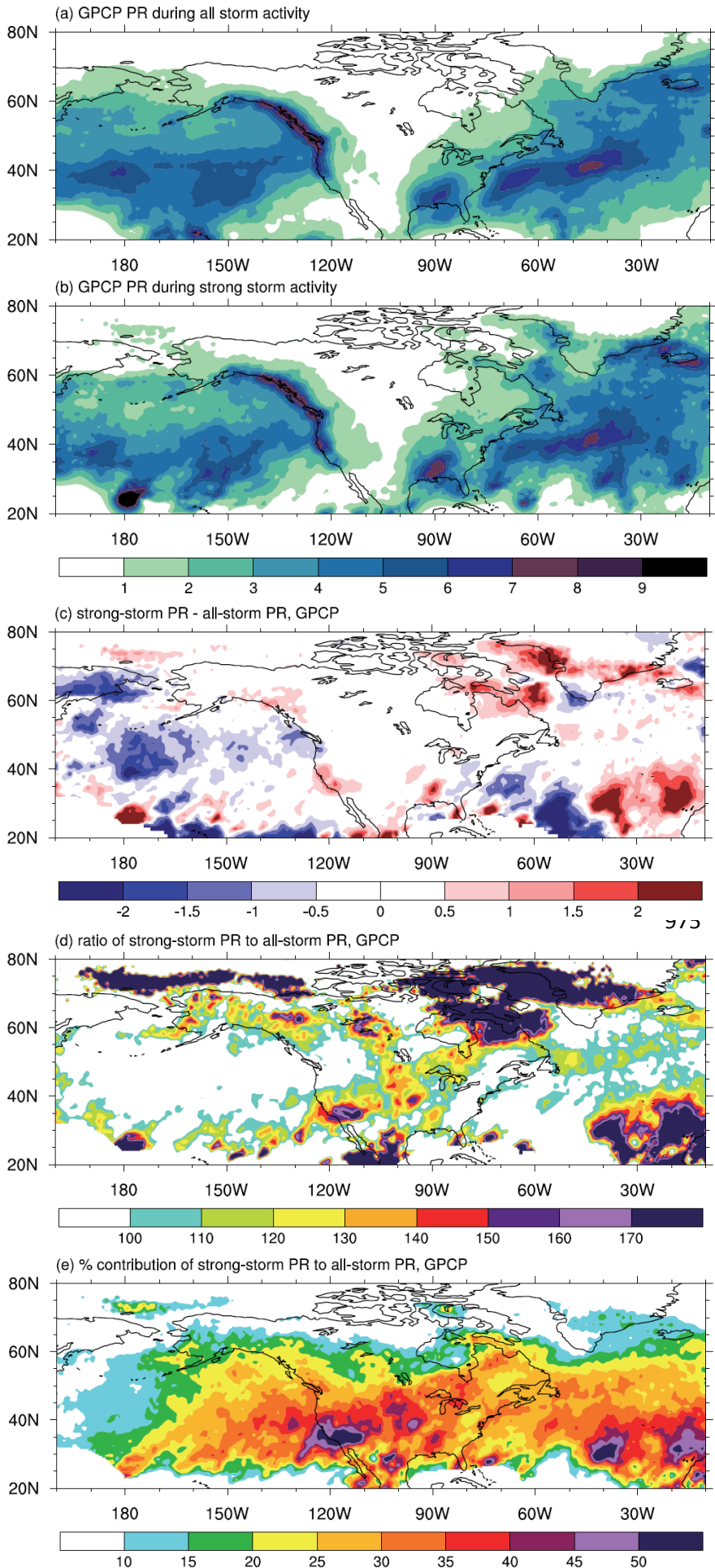
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(a) GPCP PR during all storm activity

(b) GPCP PR during strong storm activity

(c) strong-storm PR - all-storm PR, GPCP

(d) ratio of strong-storm PR to all-storm PR, GPCP

(e) % contribution of strong-storm PR to all-storm PR, GPCP

Figure 12: As in Fig. 10 but for GPCP precipitation for 1999-2010.

51

(a) GPCP all-storm PR exceeding the 98th percentile

(b) GPCP strong-storm PR exceeding the 98th percentile

(c) strong-storm PR - all-storm PR, GPCP

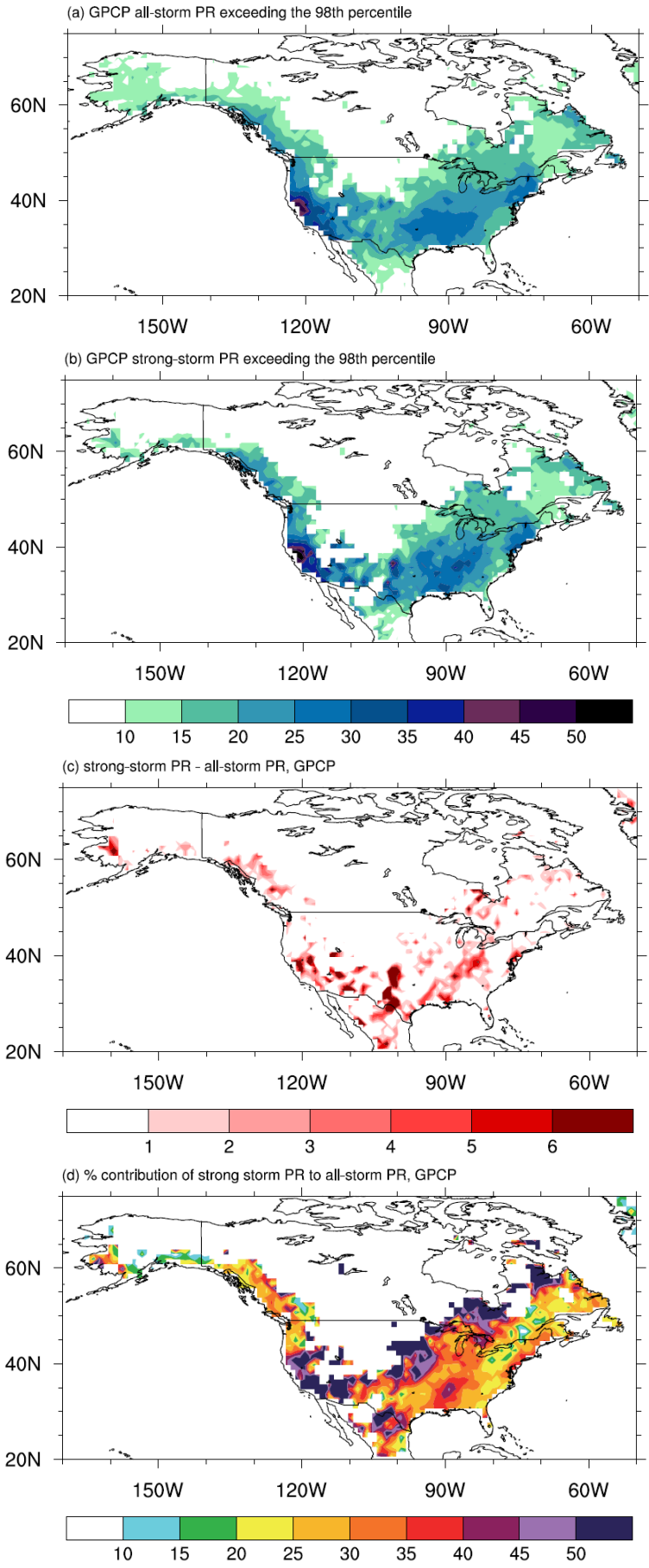(d) % contribution of strong storm PR to all-storm PR, GPCP

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