

1           **East Asian Monsoon Forcing and North Atlantic**  
2           **Subtropical High Modulation of Summer Great Plains**  
3           **Low-level Jet**

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9           **Key Points:**

- 10           • Variability of the Great Plains low-level jet is linked to the strength of the East  
11           Asian monsoon and North Atlantic subtropical high.
- 12           • North Atlantic subtropical high has a greater influence on the low-level jet and  
13           related rainfall, but the monsoon may amplify its impacts.
- 14           • Their interaction involves alignment of upper- and lower-level meridional wind anoma-  
15           lies, enhancing mid- to upper-level divergence.

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**Abstract**

Dynamic influences on summertime seasonal United States rainfall variability are not well understood. A major cause of moisture transport is the Great Plains low-level jet (LLJ). Using observations and a dry atmospheric general circulation model, this study explored the distinct and combined impacts of two prominent atmospheric teleconnections – the East Asian monsoon (EAM) and North Atlantic subtropical high (NASH) – on the Great Plains LLJ in the summer. Separately, a strong EAM and strong western NASH are linked to a strengthened LLJ and positive rainfall anomalies in the Plains/Midwest. Overall, NASH variability is more important for considering the LLJ impacts, but strong EAM events amplify western NASH-related Great Plains LLJ strengthening and associated rainfall signals. This occurs when the EAM-forced Rossby wave pattern over North America constructively interferes with low-level wind field, providing upper-level support for the LLJ and increasing mid- to upper-level divergence.

**Plain Language Summary**

Summer rainfall can greatly impact agriculture, and seasons with extreme wet or extreme dry conditions often harm human life and/or property. Therefore, it is important to understand fluctuations in seasonal rainfall and the dynamical processes involved. Moisture can be transported into the U.S. through a narrow belt of winds that peaks a few hundreds of meters above ground. We studied how the East Asian monsoon and a subtropical high system over the North Atlantic can impact this belt of winds and its related rainfall. When the North Atlantic subtropical high extends into the U.S., it strengthens the winds into the Plains and increases rainfall over the eastern U.S. A strong East Asian monsoon can amplify this response; this occurs because the monsoon triggers an atmospheric wave that crosses the North Pacific and North America, and its related flow in the upper levels can become in-phase with the belt of winds that is providing moisture to the Plains, amplifying processes that produce rainfall.

**1 Introduction**

Continental United States (CONUS) summer rainfall variability has implications for human health and the economy. Unfortunately, current subseasonal-to-seasonal forecasts for summer precipitation have relatively low skill (Becker et al., 2014; Hao et al., 2018; Jha et al., 2019; Slater et al., 2019; Malloy & Kirtman, 2020), likely because there is little to no consensus about the dominant cause(s) of low-frequency precipitation variability. Compared to winter, the warm season presents unique challenges, such as lower signal-to-noise ratios from weaker subtropical jet streams, SST anomalies, tropical convection systems, and extratropical circulation (Schubert et al., 2002; Dirmeyer et al., 2003; S. Zhou et al., 2012; Gianotti et al., 2013; Lee et al., 2009; Tian et al., 2017; Malloy & Kirtman, 2020).

The Great Plains low-level jet (LLJ) is a prominent circulation feature east of the Rocky Mountains, typically forming at nighttime just above the boundary layer between 925-850 hPa (Blackadar, 1957; Holton, 1967; T. Parish et al., 1988; Fast & McCorcle, 1990; Mitchell et al., 1995; Whiteman et al., 1997; Banta et al., 2002; Jiang et al., 2007; T. R. Parish & Oolman, 2010; Gimeno et al., 2016; Shapiro et al., 2016). Its fast-moving southerly winds act as a conveyor belt of heat and moisture to the Plains and Midwest, causing precipitation at the jet exit where low-level convergence occurs (Higgins et al., 1997; Weaver, Ruiz-Barradas, & Nigam, 2009; Pu & Dickinson, 2014; Hodges & Pu, 2019). Major pluvial events are linked to the strengthening of the Great Plains LLJ (Arritt et al., 1997; Cook et al., 2008; Feng et al., 2016) with low-level fluxes typically peaking in the mid-summer months (Weaver & Nigam, 2008); Algarra et al. (2019) found that the Great Plains LLJ contributes up to 70-90% of the moisture transport into the Plains and

65 up to 50% of the moisture transport into the Great Lakes and northeast U.S. regions in  
66 the summer.

67 There are numerous large-scale influences on Great Plains LLJ strength and vari-  
68 ability. Teleconnections, such as the Pacific-North America (PNA) pattern, are found  
69 to have strong links to LLJ strengthening (Harding & Snyder, 2015; Patricola et al., 2015;  
70 Mallakpour & Villarini, 2016; Weaver et al., 2016; Nayak & Villarini, 2017; Malloy & Kirt-  
71 man, 2020). Anomalous ridging over the northeast Pacific and anomalous troughing over  
72 western North America – characteristics of a negative PNA – promote strengthening of  
73 low-level southerlies and enhanced moisture transport over the Plains.

74 Many studies have analyzed additional trans-Pacific upper-level wave patterns and  
75 their connections to U.S. hydroclimate variability. The Asia-North America (ANA) tele-  
76 connection, an upper-level height pattern initiated by East Asian monsoon (EAM) heat-  
77 ing, has been shown to link the climate variability over Asia and North America (B. Wang  
78 et al., 2001; Lau & Weng, 2002; Zhu & Li, 2016, 2018; S. Zhao et al., 2018; Lopez et al.,  
79 2019; Malloy & Kirtman, 2022). The EAM has been shown to produce an equivalent barotropic  
80 wave train response with or without ENSO in the background state (Trenberth & Guille-  
81 mot, 1996; Lau & Weng, 2002; Zhu & Li, 2016, 2018; Lopez et al., 2019). Like the PNA,  
82 the ANA pattern is associated with an anomalous trough over western North America,  
83 promoting Great Plains LLJ strengthening. Zhu and Li (2018) found the ANA relation-  
84 ship to boreal summer rainfall variability has become stronger in recent decades, likely  
85 due to a northward shift of the monsoon system closer to the East Asian jet.

86 The North Atlantic Subtropical High (NASH) has a prominent control over large-  
87 scale circulation and the Great Plains LLJ. The NASH experiences its own variability,  
88 with its westward expansion or shift linked to Plains and/or southeast U.S. hydroclimate  
89 (Ting & Wang, 2006; L. Li et al., 2012; Pu et al., 2016; Hodges & Pu, 2019; Wei et al.,  
90 2019; Nieto Ferreira & Rickenbach, 2020). Observational analysis and an associated GCM  
91 study suggested that Indian monsoon heating may result in increasing low-level easterly  
92 wind anomalies over the North Atlantic to shift the NASH westward (Kelly & Mapes,  
93 2011, 2013). When the western ridge of the NASH intensifies, the Caribbean LLJ strength-  
94 ens, increasing the easterly transport of moisture from the subtropical Atlantic and Caribbean  
95 Sea into the Gulf of Mexico (Mestas-Nuñez et al., 2007; C. Wang, 2007; Krishnamurthy  
96 et al., 2015; García-Martínez & Bolasina, 2020; Nieto Ferreira & Rickenbach, 2020). This  
97 additionally leads to increased southerlies in the Great Plains LLJ, enhancing the mois-  
98 ture fluxes into the Plains (T. R. Parish & Oolman, 2010; Algarra et al., 2019; Z.-Z. Hu  
99 et al., 2020). Nieto Ferreira and Rickenbach (2020) determined that western NASH events  
100 are associated with 40% greater Great Plains moisture transport compared to eastern  
101 NASH events. The NASH has shifted or extended west more frequently in recent decades,  
102 and it is projected that trend will continue in a warming climate (W. Li et al., 2011; L. Li  
103 et al., 2012; Tang et al., 2017), though changes may be seasonally dependent and also  
104 controlled by poleward or equatorward shifts (W. Zhou et al., 2021). Nevertheless, un-  
105 derstanding the impacts from these changes may yield knowledge beyond seasonal or in-  
106 terannual timescales.

107 Lastly, sea surface temperature (SST) anomalies in both the Pacific and Atlantic  
108 have been linked to the summer LLJ on monthly timescales. A warm tropical and north-  
109 ern Pacific and cool north Atlantic are associated with the strengthening of the Great  
110 Plains LLJ (Ting & Wang, 1997; Weaver, Schubert, & Wang, 2009; Pegion & Kumar,  
111 2010; Q. Hu & Feng, 2012; Veres & Hu, 2013; Yu et al., 2017; Danco & Martin, 2018),  
112 though the extent to which this relationship is dynamically driven has been disputed.  
113 For example, there is a strong intraseasonal and interannual condition to the link between  
114 El Niño-Southern Oscillation (ENSO) and the Great Plains LLJ strengthening (Krishnamurthy  
115 et al., 2015; Danco & Martin, 2018). Kam et al. (2014) and Malloy and Kirtman (2020)  
116 suggest that using tropical SST forecasts for long-range rainfall prediction may be lim-  
117 iting in the summer months. Atmospheric circulation variability (internal or forced) has

118 been shown to exist in the absence of tropical forcing (A. Z. Liu et al., 1998; Ding et al.,  
 119 2011; Schubert et al., 2011; Krishnamurthy et al., 2015; Zhu & Li, 2016; O’Reilly et al.,  
 120 2018; S. Zhao et al., 2018) and may have a stronger link to Plains/Midwest hydroclimate  
 121 (Schubert et al., 2002; Ding et al., 2011; Burgman & Jang, 2015; Patricola et al., 2015;  
 122 O’Reilly et al., 2018; Malloy & Kirtman, 2020). In general, summer predictability be-  
 123 yond weather timescales has been related to the location and/or amplification of quasi-  
 124 stationary Rossby waves (Ding & Wang, 2005; Schubert et al., 2011; Beverley et al., 2019,  
 125 2021; Mariotti et al., 2020). Agrawal et al. (2021) found that monsoon-forced telecon-  
 126 nections can help explain interannual variability of the Great Plains LLJ in May; a wave  
 127 train that propagates over the U.S. can support favorable background states (i.e. enhanced  
 128 differential heating over sloping terrain) for LLJ flow. Understanding the primary forc-  
 129 ing mechanisms for these monsoon-forced planetary waves, such as from the EAM, and  
 130 how they develop over North America during the summer season where influence from  
 131 the NASH circulation is greatest, is essential. Therefore, this study will concentrate on  
 132 atmospheric teleconnections active in the June-July-August (JJA) season, particularly  
 133 the EAM’s and NASH’s relationship with the Great Plains LLJ.

134 Despite the considerable literature on the EAM and NASH and their distinct in-  
 135 fluence on CONUS rainfall variability, there is little to no exploration into how these tele-  
 136 connections interact. Because the Great Plains LLJ is a key driver of summer precip-  
 137 itation, this study will investigate the Great Plains LLJ response to the EAM forcing  
 138 and consider how the NASH modulates that response. Simple dry atmospheric general  
 139 circulation models (AGCMs) have been successful in reproducing the dynamics and vari-  
 140 ability of quasi-stationary/planetary wave activity from diabatic heating related to mon-  
 141 soons (Zhu & Li, 2016, 2018; Lopez et al., 2019; Malloy & Kirtman, 2022). We will use  
 142 a simple dry nonlinear AGCM to understand the large-scale responses and modulation  
 143 of the Great Plains LLJ on seasonal-to-interannual timescales. Because the dry AGCM  
 144 inputs surface temperature climatology, it does not simulate SST variability, effectively  
 145 isolating the atmospheric teleconnection (EAM and NASH) impacts. Section 2 will de-  
 146 scribe the datasets, details of the nonlinear AGCM and the experiments, and the rele-  
 147 vant analysis methods. Section 3 will present the results as follows: The observed responses  
 148 of the EAM and NASH and their interactions will be quantified. Then, this paper will  
 149 examine the AGCM’s EAM-forced response of the Great Plains LLJ. Finally, we will eval-  
 150 uate how NASH modulates the EAM-forced response. Section 4 will serve as a summary  
 151 and reflection of the results in the context of previous literature and future work needed.

## 152 **2 Data and Methods**

### 153 **2.1 Observational Datasets**

154 Pressure-level meridional wind, zonal wind, temperature, and geopotential height  
 155 were taken from the European Centre for Medium-Range Weather Forecasts (ECMWF)  
 156 fifth-generation reanalysis (ERA5). ERA5 atmospheric data is provided on a 0.25° lat-  
 157 itude/longitude grid (Hersbach et al., 2020). U.S. precipitation data were taken from the  
 158 CPC Unified Gauge-based Analysis, provided on a 0.25° latitude/longitude grid (Chen  
 159 et al., 2008; Xie et al., 2007). This study used the June through August monthly data  
 160 between 1979-2019 to serve as observations.

### 161 **2.2 Model and Experiments**

162 The model in this study is a dry, baroclinic, and nonlinear AGCM, i.e. it includes  
 163 the full primitive equations of divergence, vorticity, temperature and surface pressure.  
 164 It is a spectral model with Rhomboidal truncation at R42 – approximately 1.7° latitude  
 165 by 2.8° longitude – with 26 vertical levels. The vertical levels are analogous to the Com-  
 166 munity Atmospheric Model, version 4 (CAM4), which uses hybrid sigma-pressure coordi-  
 167 nate system. The AGCM is adapted from Brenner (1984) to remove moist processes.

168 Newtonian cooling is specified throughout the troposphere with enhanced damping near  
 169 the surface. Rayleigh friction is specified at the lower levels and mimic realistic land-sea  
 170 frictional contrasts to generate climatological features, such as the NASH, monsoonal  
 171 systems, and the Great Plains LLJ. Realistic topography is also an important aspect of  
 172 this model as the large-scale Great Plains LLJ requires topographical modulation of station-  
 173 ary flow (Byerle & Paegle, 2003; T. R. Parish & Oolman, 2010; Ting & Wang, 2006;  
 174 Weaver & Nigam, 2011). Versions of this dry AGCM have been used in Kirtman et al.  
 175 (2001) and is described in more detail in Malloy and Kirtman (2022). The AGCM has  
 176 also been used by He et al. (2014) to diagnose Rossby wave generation in some climate  
 177 sensitivity experiments and by Arcodia and Kirtman (2022) to examine the combined  
 178 ENSO and MJO teleconnection. This simple, idealized model is used for evaluating the  
 179 large-scale teleconnections, primarily quasi-stationary wave activity, and it exhibits simple  
 180 dry dynamic processes.

181 The surface temperature climatology for JJA is input as background state for the  
 182 model. This climatology was calculated from ERA5 data and interpolated to the model's  
 183 grid. Each experiment was integrated forward for 900 days with the JJA background state  
 184 to estimate the steady-state response for both seasonal and interannual analysis. Anal-  
 185 ysis excludes the first 100 days to assure that there is no contamination from the spin-  
 186 up period.

187 This AGCM is used for both unforced and forced experiments. The unforced ex-  
 188 periment, or control (hereby CTRL) run, is evaluated to compare climatology with ob-  
 189 servations. It is also compared to the EAM-forced runs to understand NASH modula-  
 190 tion of the Great Plains LLJ, divergence, and circulation response in the model. The strong  
 191 EAM experiment applies a constant diabatic heating via Gaussian bubble with a max-  
 192 imum of  $2 \text{ K day}^{-1}$  centered at  $30^\circ\text{N}$ ,  $120^\circ\text{E}$  and 300 hPa (see Supplementary Figure 1),  
 193 similar to Zhu and Li (2016) and identical to Malloy and Kirtman (2022). The weak EAM  
 194 experiment applies a forcing in the same location and of the same magnitude, but with  
 195 the opposite sign i.e. there is negative diabatic heating (or cooling).

### 196 2.3 Analysis Methods

197 To investigate the separate and combined roles of the EAM and NASH in both ob-  
 198 servations and the AGCM, we calculated difference composites of 900-hPa meridional  
 199 wind (V900) anomalies, 250-hPa geopotential height (Z250) anomalies, and rainfall anoma-  
 200 lies. This means that anomalies are averaged for upper tercile events, and then subtracted  
 201 from anomalies averaged from lower tercile events. We chose a composite analysis to high-  
 202 light any nonlinearities in responses as weak and strong events may not yield equal and  
 203 opposite LLJ anomalies. The EAM index is defined by 200-hPa zonal wind (U200) cir-  
 204 culation as described in G. Zhao et al. (2015):  $U200(2.5-10^\circ\text{N}, 105-140^\circ\text{E}) - U200(17.5-$   
 205  $22.5^\circ\text{N}, 105-140^\circ\text{E}) + U200(30-37.5^\circ\text{N}, 105-140^\circ\text{E})$ , where U200 is averaged anomalies within  
 206 the domain in the parentheses. The western NASH index is defined as follows:  $Z850(15-$   
 207  $28^\circ\text{N}, 50-85^\circ\text{W})$ . A variation of this intensity index was used by L. Li et al. (2012) and  
 208 Nieto Ferreira and Rickenbach (2020) in evaluating Z850 anomaly fields associated with  
 209 the Great Plains LLJ strengthening, but this index highlights northern NASH variabil-  
 210 ity, which impacts Plains/Midwest rainfall variability to a greater extent. Overall, the  
 211 index distinguishes between strong western NASH events, with the western ridge over  
 212 North America, and weak western NASH events, with the western ridge remaining over  
 213 the Atlantic (cf. Figure 1g-i, purple vs. green contours). All indices are standardized be-  
 214 fore anomalies are composited. We also composited the 1560 geopotential meter (1560-  
 215 gpm) lines for observations corresponding to the strong and weak events to signify the  
 216 NASH extent (W. Li et al., 2011) for the samples.

217 In addition, these composites are organized by a secondary condition, e.g. west-  
 218 ern NASH-related anomalies are further differentiated by strong (upper tercile) or weak

219 (lower tercile) EAM events before averaging. To assess the significance of these differ-  
 220 ence composites, we performed a two-sided Wilcoxon rank-sum test. This test is preferred  
 221 because it does not assume a Gaussian distribution, but it compares two samples' pop-  
 222 ulation mean ranks by considering if their distributions are the same.

223 To understand potential processes associated with these difference composites in  
 224 observations and CTRL experiment, we included composites of meridional wind anomaly  
 225 profiles averaged between 25-30°N, the latitude where the Great Plains LLJ and its re-  
 226 lated V900 anomalies are located. These composites separate by weak/strong western  
 227 NASH and weak/strong EAM. This aids in visualizing the interactions in the vertical.

228 Finally, we assessed NASH's influence on the EAM-forced responses in the dry AGCM  
 229 using difference of the composites, i.e. strong – weak EAM response during strong west-  
 230 ern NASH events minus strong – weak EAM response during weak western NASH events.  
 231 This determines whether the dry AGCM can simulate the correct tendency of the response  
 232 by NASH modulation. Anomalies are calculated by subtracting the climatology from the  
 233 CTRL experiment, and weak/strong western NASH events are based on the lower/upper  
 234 quintile thresholds calculated from the CTRL experiment.

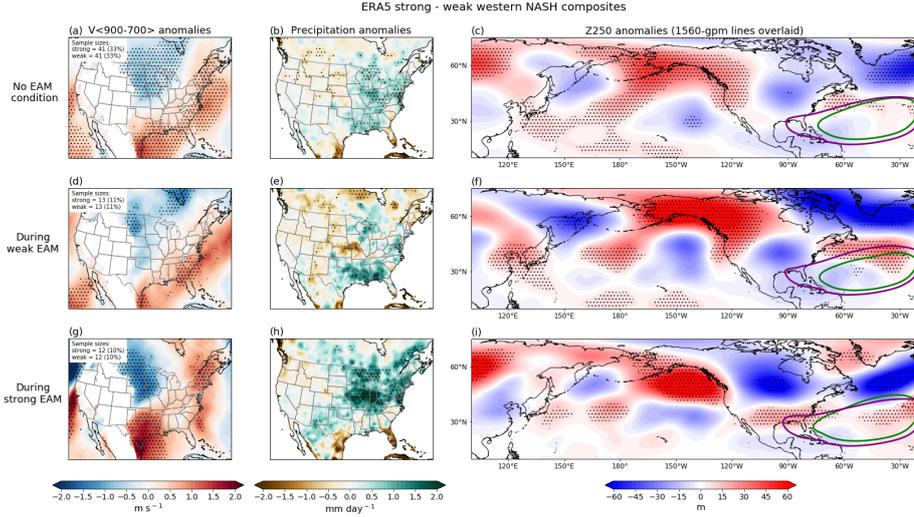
### 235 3 Results

#### 236 3.1 Observed Conditional Composite Analysis

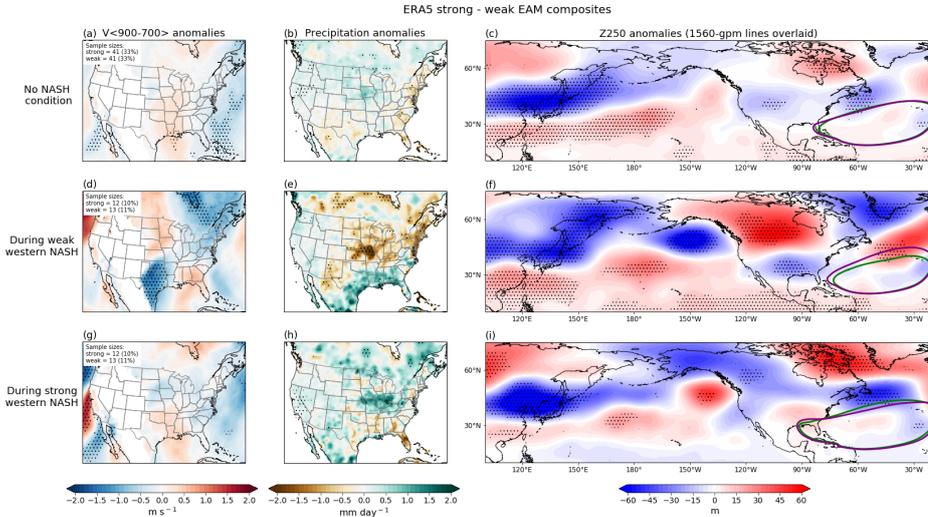
237 Figure 1 and Figure 2 decompose the separate and combined influences of the EAM  
 238 and NASH teleconnections in ERA5. The strong – weak western NASH difference com-  
 239 posites in Figure 1a-c indicate that a western NASH is related to a 1-1.5 m s<sup>-1</sup> strength-  
 240 ening of the Great Plains LLJ (Fig. 1a) and up to 1 mm day<sup>-1</sup> rainfall anomalies over  
 241 most of the eastern U.S. (Fig. 1b). The strong western NASH events (purple contour)  
 242 correspond with a 1560-gpm line that extends far into the Gulf States, consistent with  
 243 previous literature that connects west NASH extensions or shifts with amplified LLJ-  
 244 related rainfall (W. Li et al., 2011; L. Li et al., 2012). There is also an anomalous ridge-  
 245 trough pattern oriented west-east over North America (Fig. 1c).

246 When considering the strength of the EAM, the anomalous circulation and rain-  
 247 fall discussed above varies. Difference composites evaluated during weak EAM events  
 248 (Figure 1d-f) show a southward-shifted Great Plains LLJ that does not extend far into  
 249 CONUS (Fig. 1d). Rainfall anomalies of ~1.5-2 mm day<sup>-1</sup> are found over the Gulf States  
 250 only, with dry anomalies over parts of the Plains/Midwest (Fig. 1e). In contrast, dur-  
 251 ing a strong EAM (Fig. 1g-i), the Great Plains LLJ strengthening is greater (>2 m s<sup>-1</sup>)  
 252 and penetrates further into the U.S. (Fig. 1g). This is related to more extreme wet anoma-  
 253 lies (>2 mm day<sup>-1</sup>) stretching from the Plains to the Northeast U.S. The NASH-related  
 254 Z250 anomalies are different between weak and strong EAM events (Fig. 1f,i), partic-  
 255 ularly over East Asia, North America, and the North Atlantic. The north-south orien-  
 256 tation of anomalous trough-ridge pattern over CONUS during strong EAM events sig-  
 257 nals a negative PNA and enhanced meridional transport (Harding & Snyder, 2015; Mal-  
 258 loy & Kirtman, 2020).

259 We considered the reverse analysis as well by taking strong – weak EAM difference  
 260 composites of V900 anomalies, further separated into weak or strong western NASH events,  
 261 as seen in Figure 2. A strengthened EAM is associated with a ~0.5 m s<sup>-1</sup> strengthen-  
 262 ing of the Great Plains LLJ (Fig. 2a), though is further east from the Rockies than the  
 263 climatological Great Plains LLJ location and the NASH-related LLJ strengthening. Rain-  
 264 fall anomalies are modest – up to 0.75 mm day<sup>-1</sup> – in the northern Plains and into Canada  
 265 (Fig. 2b). An anomalous ridge is stretched over the North Pacific at around 30°N with  
 266 an upper-level wave pattern emanating northward over East Asia (Fig. 2c). In addition,  
 267 there is a general southwest-northeast pattern of an anomalous trough-ridge over North  
 268 America, and an anomalous trough off the coast of Northeast U.S.



**Figure 1.** ERA5 strong – weak western NASH difference composites (a-c) with no EAM condition considered, (d-f) only during weak EAM events, and (g-i) only during strong EAM events. Difference composites of (a,d,g) V900 anomalies, (d,e,h) CPC gauge-based precipitation anomalies, and (c,f,i) Z250 anomalies, with purple and green contours denoting the 1560-gpm line for strong and weak composites, respectively. Sample sizes for the composites and the percentage of total events the samples represent are annotated on top left of each row. Stippling indicates anomalies significant at 90% confidence level based on the Wilcoxin rank-sum test.

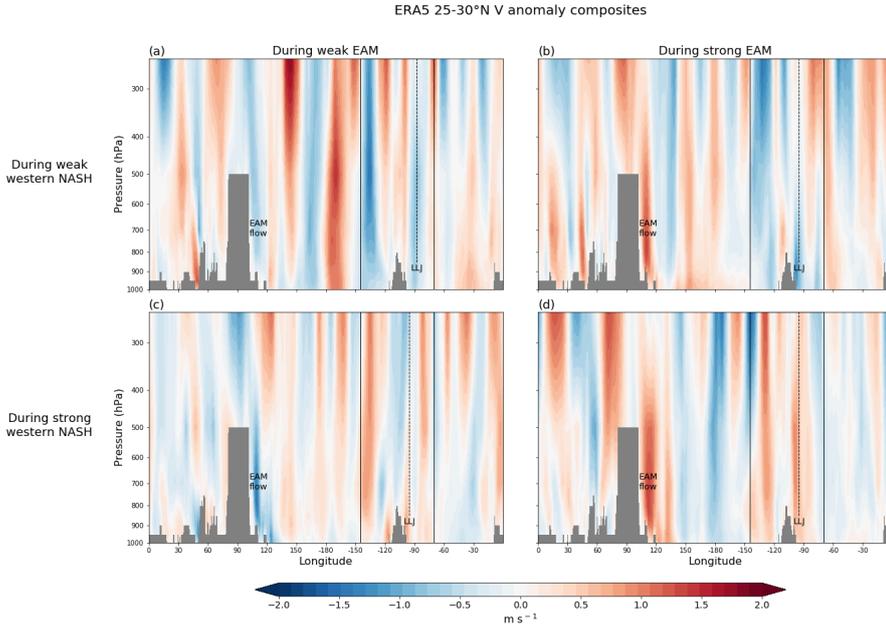


**Figure 2.** Similar format as Fig. 1, but ERA5 strong – weak EAM difference composites (a-c) with no NASH condition considered, (d-f) only during weak western NASH events, and (g-i) only during strong western NASH events.

269 Difference composites taken during weak western NASH events (Fig. 2d-f) reveal  
 270  $\sim 1.5 \text{ m s}^{-1}$  weakening of the Great Plains LLJ (Fig. 2d) and  $\sim 2 \text{ mm day}^{-1}$  dry anomalies  
 271 over the Plains, Midwest, and Northeast U.S. (Fig. 2e). During strong western NASH

272 events (Figure 2g-i), the Great Plains LLJ strengthening is weakly positive but not sta-  
 273 tistically significant (Fig. 2g), and there are  $\sim 1\text{-}1.5\text{ mm day}^{-1}$  wet anomalies (Fig. 2h).  
 274 The greatest differences in upper-level wave pattern are seen over the eastern North Pa-  
 275 cific and North America, with opposite patterns depending on western NASH strength.  
 276 This suggests that NASH exerts the primary influence over the Great Plains LLJ regard-  
 277 less of the EAM strength. In addition, because the anomalies are only statistically sig-  
 278 nificant during weak western NASH events, EAM-related wave patterns may destructively  
 279 interfere with strong western NASH-related wave patterns.

280 To further understand processes between the strong and weak events, Figure 3 shows  
 281 the vertical profile of meridional wind anomalies averaged between  $25\text{-}30^\circ\text{N}$ . Rows dif-  
 282 ferentiate between NASH strength, and columns differentiate between EAM strength.  
 283 The EAM-related flow can be discerned east of the Himalayas ( $100\text{-}120^\circ\text{E}$ ) by the northerlies  
 284 (Fig. 3a,c) or southerlies (Fig. 3b,d), which signals whether there is low-level diver-  
 285 gence or convergence over the EAM region, respectively. The Great Plains LLJ is found  
 286 between  $-100$  and  $-90^\circ\text{W}$ , with northerlies coinciding with a weak western NASH (Fig.  
 287 3a,b) and southerlies coinciding with a strong western NASH (Fig. 3c,d). During weak  
 288 EAM and weak western NASH events (Fig. 3a), as well as during strong EAM and strong  
 289 western NASH events (Fig. 3d), the LLJ-related winds are of the same sign as the upper-  
 290 level flow. This suggests that when the EAM and western NASH are both weak or strong,  
 291 their related circulation patterns are in constructive interference, i.e. the low- and upper-  
 292 level flow are in alignment to promote enhanced precipitation patterns. This alignment  
 293 of the meridional wind anomalies does not occur when the EAM is strong and the west-  
 294 ern NASH is weak, or vice versa (Fig. 3b,c).

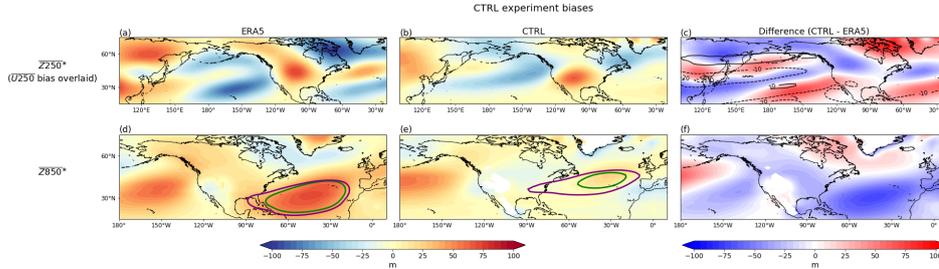


**Figure 3.** ERA5 composites of V anomaly vertical profiles averaged between  $25\text{-}30^\circ\text{N}$  during (a) weak western NASH and weak EAM events, (b) weak western NASH and strong EAM events, (c) strong western NASH and weak EAM events, and (d) strong western NASH and strong EAM events. Each panel annotates the approximate location of EAM-related flow and the Great Plains LLJ, and a thin vertical dotted line from LLJ is displayed to visualize upper-level support (or lack thereof).

295 While the idea that monsoon-forced teleconnections can influence low-level flow and  
 296 related precipitation is supported by previous literature (Harding & Snyder, 2015; Mal-  
 297 lakpour & Villarini, 2016; Agrawal et al., 2021; Malloy & Kirtman, 2022), we recognize  
 298 the small sample sizes of the difference composites from the observational dataset. There-  
 299 fore, in the next part of this study, we explore whether a simple dry AGCM can repro-  
 300 duce this interference between the EAM and NASH in influencing the Great Plains LLJ.  
 301 In addition, it might be possible that a common driver, like ENSO, is modulating EAM-  
 302 NASH interactions. Though we inspected the months that went into each composite and  
 303 did not note any composites or phase of EAM/NASH that heavily favored an ENSO phase  
 304 (see Supplementary Table 1), it would be advantageous to use the dry AGCM since it  
 305 does not simulate SST variability and hence we can isolate the atmospheric influence.

### 3.2 Control Experiment Climatology and Biases

307 Before analyzing the dry AGCM responses, we evaluated the climatological biases  
 308 of the model and its ability to produce realistic dynamic responses (e.g. quasi-stationary  
 309 Rossby waves). Zonally-asymmetric components (represented by  $*$ ) of time-mean circula-  
 310 tion (represented by  $-$ ) – also known as stationary waves – are useful for understand-  
 311 ing the production and maintenance of Rossby waves. Seasonally, stationary waves de-  
 312 scribe preferred locations of meridional fluxes of heat and moisture, affecting hydroclima-  
 313 te. We compared the stationary waves in ERA5 and the CTRL experiment (no heat-  
 314 ing forcing) from the dry AGCM (Figure 4). In observations,  $\overline{Z250}^*$  generally features  
 315 high pressure over the continents and low pressures over the ocean basins in the mid-  
 316 latitudes and subtropics at the edge of the East Asian or North Atlantic jet (Fig. 4a).  
 317 The CTRL experiment exhibits similar patterns (Fig. 4b) but with biases over the Pa-  
 318 cific and Atlantic (Fig. 4c). The bands of low pressure are higher in latitude over the  
 319 Pacific, and the Atlantic is missing a band of low pressure at the subtropics. This has  
 320 implications for the location of jet streams; the CTRL experiment  $\overline{U250}$  bias (overlaid  
 321 on Fig. 4c) indicates a jet stream shifted northward. Overall, the model captures basic  
 322 characteristics of upper-level circulation, but these biases are important for understand-  
 323 ing the production of Rossby wave responses in the forcing experiments.

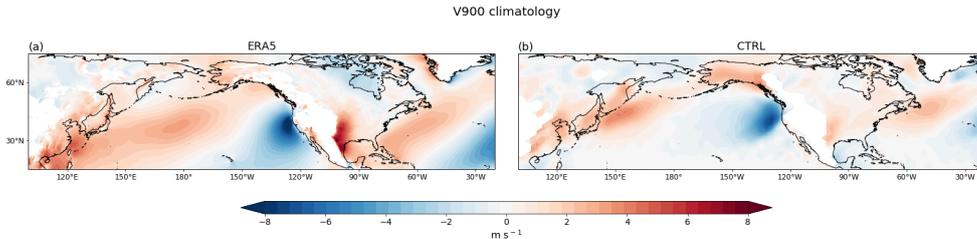


**Figure 4.** (a-c) Zonally-asymmetric component of the Z250 climatology for (a) ERA5, (b) CTRL experiment, and (c) the subtraction difference between CTRL experiment and ERA5, with the U250 climatological bias overlaid (black contours). (d-f) Same as top row, but for Z850. The purple and green contours denote the 1560-gpm line for the strong and weak NASH events, respectively, for both (d) ERA5 and (e) CTRL experiment.

324  $\overline{Z850}^*$  from observations presents high pressure systems over the ocean basins (Fig.  
 325 4d), which coincides with the climatological location of subtropical highs (e.g. NASH).  
 326 These ridges generally appear in the CTRL experiment (Fig. 4e), but the NASH is weaker  
 327 and further north. These biases may have implications for discerning NASH influences,  
 328 e.g. related anomalies that are higher in latitude than observations. Nevertheless, NASH

329 variability is simulated in this model, as seen by comparing the composited positive and  
 330 negative western NASH events between ERA5 and CTRL experiment (purple and green  
 331 contours in Fig. 4d,e). The dry AGCM simulates strong western NASH events with west-  
 332 ern ridge extensions over CONUS and weak western NASH events with western ridge  
 333 extensions that remain over the Atlantic, though the NASH extents are generally fur-  
 334 ther north and exhibit greater variability between the weak and strong events. Overall,  
 335 the basic NASH circulation and variability and its connection to the Great Plains LLJ  
 336 is represented.

337 V900 climatology, which indicates the Great Plains LLJ climatology, can be com-  
 338 pared in Figure 5. The ERA5 time-mean V900 shows a strong ( $\sim 8 \text{ m s}^{-1}$ ) Great Plains  
 339 LLJ feature (Fig. 5a). Despite the climatological core being about  $5^\circ$  northward from  
 340 observational estimates, the location of the Great Plains LLJ in CTRL experiment is close  
 341 to the Rockies, and general V900 circulation features over East Asia, North Pacific, and  
 342 North America are represented in the model despite a northward-shifted bias. The mag-  
 343 nitude of the climatological core is  $3 \text{ m s}^{-1}$ , which is weaker than the observations. How-  
 344 ever, the objective of the study is to analyze large-scale dynamical differences between  
 345 forced experiments, not to represent thermodynamics, diurnally-varying radiative pro-  
 346 cesses, nor mesoscale physics; therefore, this simulated Great Plains LLJ is within rea-  
 347 son given that the model has relatively coarse resolution and lacks moist processes and  
 348 associated land-atmosphere feedbacks.

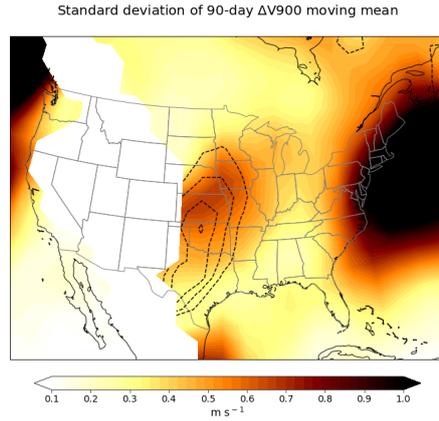


**Figure 5.** V900 climatology for (left) ERA5 and (right) CTRL experiment.

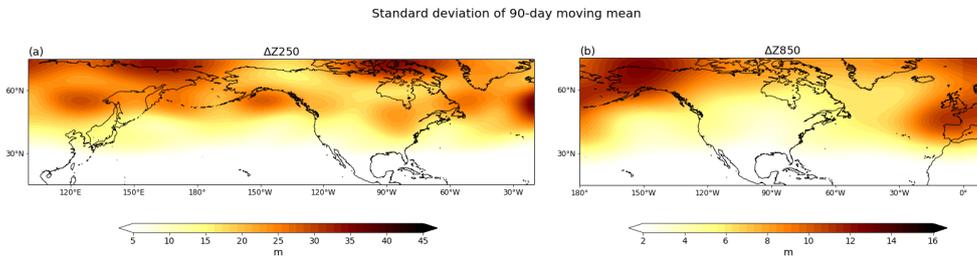
### 349 **3.3 Strong – Weak EAM-forced Experiment Analysis**

350 An advantage of this experiment setup with the dry nonlinear AGCM is that one  
 351 can assess the effect of sub-sampling 90-day (or one single season) means during the 900-  
 352 day experiment. Figure 6 demonstrates the internal variability of 90-day V900 means  
 353 for this experiment; for V900 responses, the  $\sigma$  values are relatively large on the north-  
 354 ern and southern edges of the climatological Great Plains LLJ region, and substantial  
 355 off both North American coasts. This suggests that fluctuations in V900 are primarily  
 356 at the northern edge of the climatological LLJ. The Z250 and Z850 responses indicate  
 357 relatively higher  $\sigma$  values along the approximate climatological jet stream latitude and  
 358 along the boundaries of climatological subtropical highs, respectively (Figure 7). This  
 359 likely means that fluctuations in upper and lower heights are linked to East Asian jet  
 360 variations and shifts in the subtropical highs, respectively. By dividing the time-mean  
 361 difference by this standard deviation, we assess the robustness (or statistical significance)  
 362 of the long-term response on seasonal-to-interannual timescales.

363 The EAM-forced V900 response is summarized in Figure 8, indicating a 0.5-1 m  
 364  $\text{s}^{-1}$  strengthening – a  $\sim 25\%$  magnitude increase compared to Figure 5b – in the Great  
 365 Plains LLJ. This strengthening is confined to the northern side of the jet (Fig. 8c), which  
 366 differs from the strong – weak EAM difference composite in Figure 2a. Overall, this forced



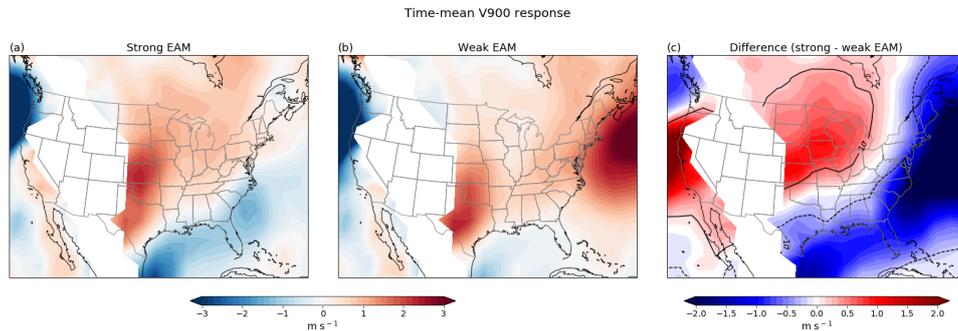
**Figure 6.** Standard deviation of the 90-day  $\Delta V900$  ( $\Delta =$  strong EAM experiment - weak EAM experiment) moving mean. Climatological Great Plains LLJ is overlaid (black dashed contours).



**Figure 7.** Standard deviation of the 90-day (left)  $\Delta Z250$  and (right)  $\Delta Z850$  moving mean.

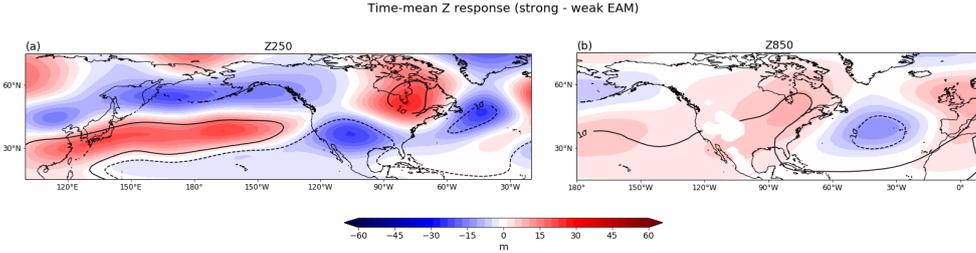
367  
368

response is considered robust on the seasonal timescale in the Great Plains and over the Gulf of Mexico, seen by the positive (negative) difference values that exceed  $1\sigma$  ( $-1\sigma$ ).



**Figure 8.** Time-mean V900 climatology for (a) strong EAM experiment and (b) weak EAM experiment. (c) Subtraction difference between strong EAM experiment and weak EAM experiment time-mean V900, with the  $1\sigma$  (solid black) and  $-1\sigma$  (dashed black) contours overlaid by dividing difference by the standard deviation of the 90-day  $\Delta V900$  moving mean.

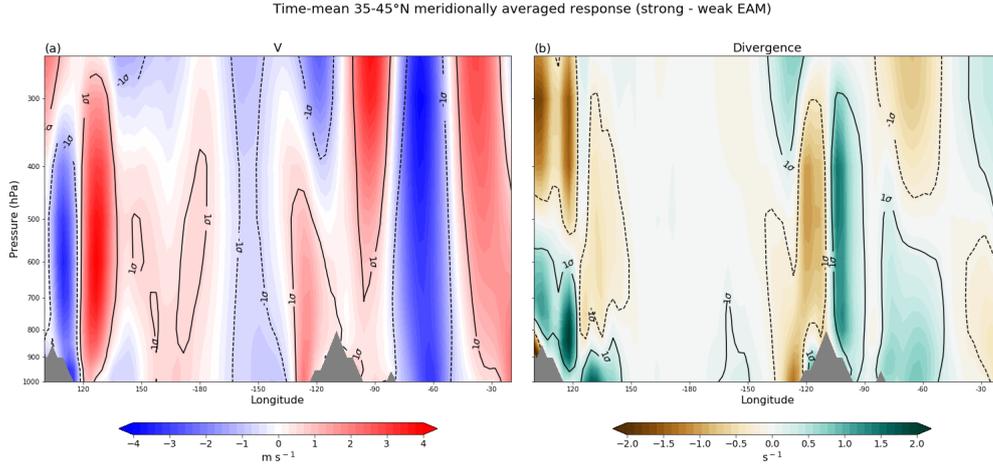
369 The EAM Z250 time-mean response shows zonally-oriented troughs and ridges that  
 370 stretch from the EAM region and over the North Pacific (Figure 9a), with an anomalous  
 371 trough-ridge pattern oriented west-east over North America, similar to the observed  
 372 pattern (cf. Figure 2c). The anomalous trough over western North America is typically  
 373 associated with Great Plains LLJ strengthening (Harding & Snyder, 2015; Mallakpour  
 374 & Villarini, 2016; Malloy & Kirtman, 2020). The EAM Z850 time-mean response (right)  
 375 presents anomalous ridging over much of the North Pacific and North America and anomalous  
 376 troughing over the mid-latitude Atlantic, which could signal an increased variability  
 377 of the NASH in the west-east direction. The Z250 and Z850 responses are mostly ro-  
 378 bust except for the high latitudes and the eastern North Pacific/Alaska region.



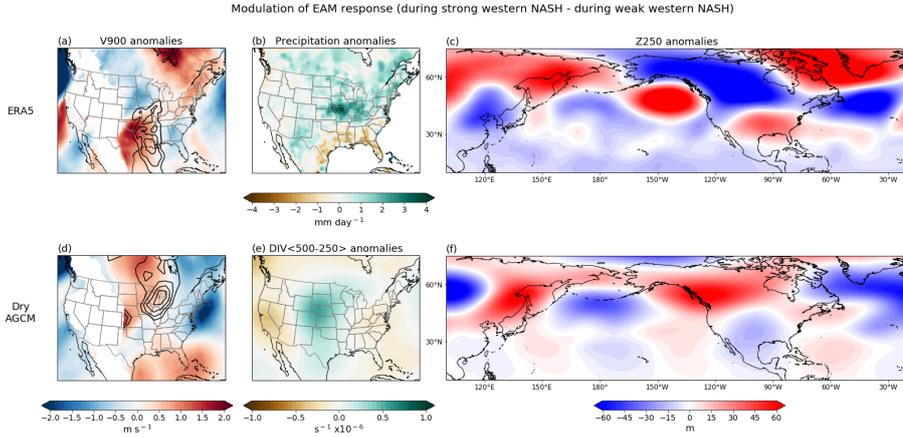
**Figure 9.** Subtraction difference between strong EAM experiment and weak EAM experiment zonally-asymmetric component of the time-mean (a) Z250 and (b) Z850. The  $1\sigma$  (solid black) and  $-1\sigma$  (dashed black) contours are overlaid as in Fig. 8.

379 To get a sense of the response in the vertical, we assessed the latitudinally aver-  
 380 aged cross-section of strong – weak EAM meridional wind ( $V$ ) and divergence response  
 381 in the general region where the downstream wave response travels ( $35\text{-}45^\circ\text{N}$ ; Figure 10).  
 382 The response is mostly equivalent barotropic except for over Gulf of Alaska/eastern North  
 383 Pacific. However, the most statistically significant  $\Delta V$  values are located over the EAM  
 384 region as well as North America (Fig. 10a), including the upper-level trough and ridge  
 385 from Figure 9. This corresponds to the anomalous divergence on the leeside of the Rock-  
 386 ies (Fig. 10b). Despite the robust differences in this region, there is still substantial in-  
 387 ternal variability over the mid-latitude Pacific and/or the upper levels.

388 Lastly, we evaluated the influence of NASH on the EAM responses, visualized by  
 389 taking the strong – weak EAM responses during strong western NASH events and sub-  
 390 tracting by the strong – weak EAM responses during weak western NASH events (Figure  
 391 11), done for both observations (Fig. 11a-c) and the dry AGCM (Fig. 11d-f). The  
 392 climatological biases of the dry AGCM are apparent, with Great Plains LLJ strength-  
 393 ening  $10^\circ$  northward from the observational strengthening (Fig. 11a,d). However, by con-  
 394 sidering these biases and comparing the NASH-modulated strong – weak EAM response  
 395 (shaded contours) with the original strong – weak EAM response (no NASH considered,  
 396 solid black contours), it is evident that strong western NASH modulation is compar-  
 397 able between observations and the dry AGCM, i.e. a strong western NASH amplifies the  
 398 Great Plains LLJ strengthening signal, especially on the side closest to the Rockies. The  
 399 dry AGCM generally simulates enhanced 500-250-hPa layer-averaged divergence in the  
 400 Plains associated with the enhanced precipitation anomalies from observations (Fig. 11b,e).  
 401 This suggests that dry dynamics in the AGCM may be sufficient to produce basic NASH-  
 402 related modulation of EAM-forced patterns, such that the signs of the response are cor-  
 403 rect (see Supplementary Figure 2 for full strong – weak EAM response separated by west-  
 404 ern NASH strength as in Figure 2). Z250 patterns outside North America compare well  
 405 to observations, but there are discrepancies in the dry AGCM representation of NASH  
 406 modulation of Z250 over North America (Fig. 11c,f) that may limit its representation.



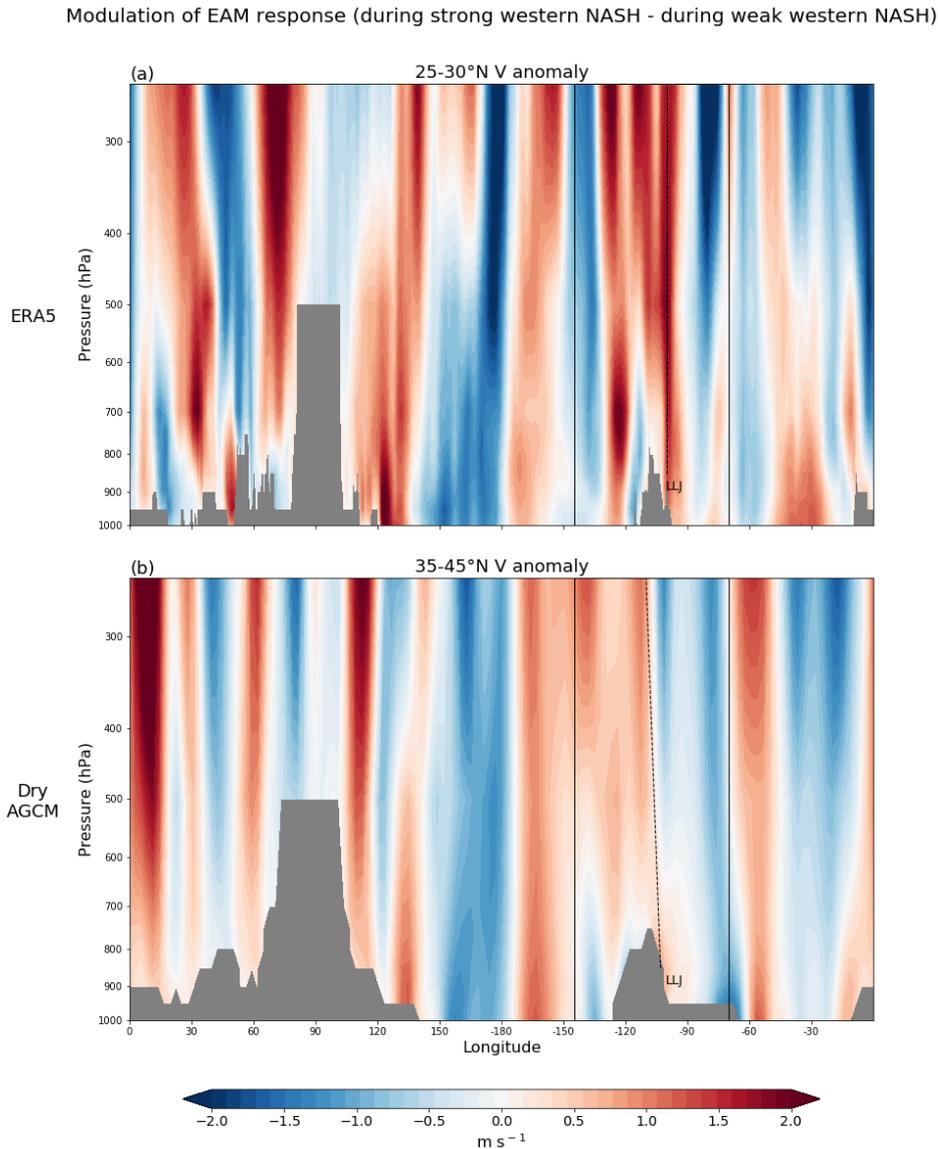
**Figure 10.** Subtraction difference between strong EAM experiment and weak EAM experiment 35-45°N meridionally averaged time-mean profile of (a) V and (b) divergence. The  $1\sigma$  (solid black) and  $-1\sigma$  (dashed black) contours are overlaid as in Fig. 8.



**Figure 11.** Modulation of EAM response by NASH: (a-c) Subtraction difference between ERA5 strong – weak EAM composites during strong western NASH and weak western NASH, i.e. Fig. 2g-i minus Fig. 2d-f. (d-f) Subtraction difference between strong EAM experiment – weak EAM experiment, i.e. strong – weak EAM response during strong western NASH minus strong – weak EAM response during weak western NASH, with 500-250-hPa layer-averaged divergence anomalies instead of precipitation anomalies. Strong – weak EAM V900 anomalies without NASH condition in Great Plains LLJ region are overlaid for left column for reference.

407 NASH modulation is further demonstrated by taking vertical profiles of meridional  
 408 wind where the Great Plains LLJ strengthening occurs (averaged 25-30°N for observa-  
 409 tions and averaged 35-45°N for dry AGCM; Figure 12). NASH modulation of circula-  
 410 tion is notably similar to observations and the dry AGCM except over North Atlantic.  
 411 Over the region of interest that affects the Great Plains LLJ, the dry AGCM presents  
 412 alignment of positive meridional wind values from the low to upper levels (Fig. 12b), though  
 413 not as vertically stacked as presented in observations (Fig. 12a) or Figure 3. Overall, the  
 414 dry AGCM simulates NASH modulation of EAM-forced responses to a reasonable de-

415 gree, including the amplification of Great Plains LLJ strengthening and related diver-  
 416 gence during strong EAM and strong western NASH events.



**Figure 12.** Modulation of vertical profile of EAM response by NASH: (a) Subtraction difference between ERA5 strong – weak EAM 25-30°N V composites during strong western NASH and weak western NASH. (b) Subtraction difference between strong EAM experiment – weak EAM experiment, i.e. strong – weak EAM 35-45°N V response during strong western NASH minus strong – weak EAM 35-45°N V response during weak western NASH. Each panel annotates the approximate location of the Great Plains LLJ, and a thin vertical dotted line from LLJ is displayed to visualize upper-level support.

## 417 4 Summary and Discussion

418 Seasonal forecasts of CONUS precipitation during the summer have relatively low  
 419 skill, and there is little consensus on the driving causes of rainfall variability on this timescale.  
 420 We suggest that examining large-scale Great Plains LLJ responses in observations and  
 421 a dry nonlinear AGCM will aid in discerning dynamic causes and variability of pluvial  
 422 events. First, we compared observational analysis of the NASH and EAM teleconnec-  
 423 tions and their interactions. Then we analyzed and compared Great Plains LLJ responses  
 424 from EAM experiments in a dry AGCM and explored whether NASH modulation of EAM  
 425 circulation responses can be reproduced with simple dry dynamics.

426 Results from the ERA5 conditional difference composites (Figure 1 and Figure 2)  
 427 suggested that the strength of the western NASH or EAM matters when considering Great  
 428 Plains LLJ impacts. Strong western NASH-related Great Plains LLJ strengthening and  
 429 associated wet anomalies were greater during strong EAM events. However, EAM-related  
 430 Great Plains LLJ responses were more dependent on the NASH location: during weak  
 431 western NASH events, the strong – weak EAM response is a weakened Great Plains LLJ,  
 432 and the LLJ response during strong western NASH events is not statistically significant.  
 433 Profiles of meridional wind anomalies revealed that strong (weak) EAM and strong (weak)  
 434 western NASH events were linked to in-phase lower- and upper-level circulation patterns,  
 435 providing enhanced upper-level support for the Great Plains LLJ (Figure 3).

436 The strong – weak EAM responses were largely captured by the dry AGCM, in-  
 437 cluding an elongated wave structure over the North Pacific and anomalous trough over  
 438 western North America (Figure 9) comparable to observations (cf. Figure 2c). This pro-  
 439 moted robust Great Plains LLJ strengthening (Figure 8). In addition, the dry AGCM  
 440 simulated the amplification of the EAM-forced LLJ and mid- to upper-level divergence  
 441 during a strong western NASH due to constructive interference of low- and upper-level  
 442 wind patterns (Figures 10 and 11), shedding light on the major dynamic causes of Great  
 443 Plains LLJ strengthening and its impacts.

444 Despite the AGCM capturing many of the dynamical processes behind EAM re-  
 445 sponses and NASH modulation, there were climatological biases in the AGCM that help  
 446 explain some of the discrepancies between observations and the model’s EAM-NASH-  
 447 LLJ relationships. For example, in the upper levels, the model had a northward-shifted  
 448 jet stream corresponding to increased horizontal height gradients further north (cf. Fig-  
 449 ure 4). Accordingly, the AGCM’s Great Plains LLJ climatological core (Figure 5) and  
 450 anomalies as well as the NASH were shifted northward. Our results complement previ-  
 451 ous research that found that the inaccurate location and strength of large-scale atmo-  
 452 spheric features, such as the jet stream and subtropical high systems, can negatively im-  
 453 pact long-range forecast skill (Y. Liu et al., 2019; O’Reilly et al., 2018) or change the as-  
 454 sociated primary rainfall patterns (W. Zhou et al., 2021). Biases or discrepancies between  
 455 the observations and AGCM could also be from processes not represented in the model,  
 456 like SST variability, land-atmosphere feedbacks, or moisture processes. For example, NASH  
 457 modulation of Z250 patterns over North America was not as well represented in the dry  
 458 AGCM as the rest of the domain (cf. Figure 11); this may indicate that ENSO variabil-  
 459 ity is important to simulate NASH modulation over the continent (Malloy & Kirtman,  
 460 2022) or soil moisture-circulation feedbacks (Dirmeyer et al., 2003; Koster et al., 2006;  
 461 Jong et al., 2021).

462 Previous studies have suggested that monsoon forcing is related to the circumglobal  
 463 teleconnection (CGT), a prominent mode of upper-level height variability in the sum-  
 464 mer (Ding & Wang, 2005; Ding et al., 2011; S. Zhao et al., 2018). Typically, the CGT  
 465 wavenumber-5 pattern is maintained by Indian monsoon heating, but F. Zhou et al. (2020)  
 466 suggested the EAM influences the CGT. Agrawal et al. (2021) suggested that the CGT  
 467 greatly influences May Great Plains LLJ activity. The study found that the CGT is dy-  
 468 namically linked to both coupled and uncoupled LLJ via an enhanced geostrophic flow

469 from the upper-level wave pattern modulation. Additionally, Indian monsoon heating  
 470 may relate to NASH shifts (Kelly & Mapes, 2011, 2013). While our results show sim-  
 471 ilar features that relate the summer EAM to the CGT (F. Zhou et al., 2020) as well as  
 472 the summer CGT to the Great Plains LLJ, it is beyond the scope of this study to di-  
 473 agnose and disentangle true causal relationships between the EAM, NASH, Indian mon-  
 474 soon, and CGT. Future work will be needed to understand these inter-relationships and  
 475 how they contribute to rainfall variability over Asia, North America, and Europe.

476 A future study should expand on NASH’s role by forcing vorticity anomalies over  
 477 the western NASH region with the AGCM or investigating other sources of North At-  
 478 lantic Rossby wave activity, e.g. NAO (Weaver & Nigam, 2008). In addition, the sea-  
 479 sonal transition from early summer to late summer may also change the relationships  
 480 between the NASH, EAM, and Great Plains LLJ. Simple AGCMs have the potential to  
 481 isolate circulation responses from distinct forcing and evaluate the predictability of sum-  
 482 mer hydroclimate features. This research serves as a preliminary step for understand-  
 483 ing more complex models and assessing the predictability of atmospheric dynamics in  
 484 the summer on the more “elusive” long-range timescale.

## 485 5 Open Research

### 486 5.1 Data Availability Statement

487 All data in this study is available online. ERA5 data can be accessed through their  
 488 website <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>  
 489 (Hersbach et al., 2020). The CPC Global Unified Gauge-based Analysis data was pro-  
 490 vided by the NOAA PSL, Boulder, Colorado, USA, from their website at [https://psl](https://psl.noaa.gov)  
 491 [.noaa.gov](https://psl.noaa.gov) (Chen et al., 2008; Xie et al., 2007). The dry AGCM and the model data from  
 492 the article are available upon request.

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 496 and DOE Grant DE-SC0019433.

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