



Monsoon low pressure system like variability in an idealized moist model

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ABSTRACT

12 In this paper, it is shown that westward-propagating monsoon-low-pressure-
13 system-like disturbances in the South Asian monsoon region can be simulated
14 in an idealized moist general circulation model through the addition of a sim-
15 plified parameterization of land. Land is parameterized as having one-tenth
16 the heat capacity of the surrounding slab ocean, with evaporation limited by
17 a bucket hydrology model. In this model, the prominent topography of the
18 Tibetan Plateau does not appear to be necessary for these storm systems to
19 form or propagate; therefore focus is placed on the simulation with land but
20 no topography.

21 The properties of the simulated storms are elucidated using regression anal-
22 ysis and compared to results from composites of storms from comprehensive
23 GCMs in prior literature and reanalysis. The storms share a similar vertical
24 profile in anomalous Ertel potential vorticity to those in reanalysis. Propa-
25 gation, however, does not seem to be strongly dictated by beta-drift. Rather,
26 it seems to be more closely consistent with linear moisture vortex instability
27 theory, with the exception of the importance of the vertical advection term
28 in the Ertel potential vorticity budget toward the growth and maintenance of
29 disturbances. The results presented here suggest that a simplified GCM con-
30 figuration might be able to be used to gain a clearer understanding of the
31 sensitivity of monsoon low pressure systems to changes in the mean state cli-
32 mate.

1. Introduction

South Asia has a monsoonal climate. It receives 50 % to 70 % of its annual precipitation during the months of June, July, and August (Neelin 2007). During these months, moist static energy is abundant, fueling monsoon low pressure systems (MLPSs) which originate in the Bay of Bengal and propagate westward against the direction of the prevailing mean low-level winds, across India at speeds of around 4 m s^{-1} (Adames and Ming 2018a). Studies in recent years have attributed between 50% and 60% of monsoon season rainfall in central India to these lows (Hurley and Boos 2015; Praveen et al. 2015; Hunt and Fletcher 2019). For that reason, understanding what influences the propagation and structure of these transient phenomena is important for understanding what controls precipitation during the summer in South Asia.

The growth, propagation, and structure of these low pressure systems has been an area of research for several decades, dating back to Godbole (1977) and references therein. In recent years effort has been made by multiple independent research groups to compile detailed track information for monsoonal disturbances (Hurley and Boos 2015; Hunt et al. 2016a). This effort has led to new insights resulting from rigorous analysis of the composite properties of these storms (Hurley and Boos 2015; Boos et al. 2015; Ditchek et al. 2016; Hunt et al. 2016a,b; Cohen and Boos 2016; Sandeep et al. 2018). In particular, early theoretical attempts to explain the growth and propagation of monsoon depressions in terms of barotropic (Shukla 1977; Lindzen et al. 1983), baroclinic (Mishra and Salvekar 1980; Mak 1983; Moorthi and Arakawa 1985), or combined barotropic-baroclinic (Krishnamurti et al. 1976; Shukla 1978) instability mechanisms have recently been challenged by a number of alternative ideas.

An example where these early ideas were challenged is the study by Cohen and Boos (2016). They investigated composites of observed monsoon depressions in reanalysis and compared them

56 with the canonical example of moist baroclinic instability: diabatic Rossby waves in the mid-
57 latitudes. They found that in monsoon depressions, anomalies in Ertel potential vorticity do not
58 tilt against the mean vertical wind shear as they do in diabatic Rossby waves, which they argue is
59 evidence against moist baroclinic instability operating as a mechanism in fueling the growth of the
60 disturbances. In their paper Cohen and Boos (2016) also invoke results from Krishnamurti et al.
61 (2013) to argue that barotropic instability plays a minor, if any, role in the development of MLPSs.
62 In Krishnamurti et al. (2013) it was found that kinetic energy from the eddies in observed MLPSs
63 was transferred to the mean zonal flow, counter to what occurs in barotropically unstable flows.

64 In the last five years, four, possibly overlapping, alternative explanations for monsoonal dis-
65 turbance propagation have been proposed. The first is that monsoon depressions might be better
66 described as tropical-cyclone-like features propagating via adiabatic beta drift (Boos et al. 2015),
67 though perhaps without as strong a dependence on surface fluxes, which have been shown to be
68 important for tropical cyclones (Muller and Romps 2018). Another possible explanation, proposed
69 in Hunt and Parker (2016), is that the Himalayan mountains may act as a rigid northern meridional
70 boundary in the lower troposphere, leading to westward propagation of a cyclonic vortex to the
71 south via an effective mirror-image vortex. Adames and Ming (2018b) develop a linear theory
72 for monsoonal disturbances within a mid-latitude moisture-mode like framework, where the insta-
73 bility depends necessarily on the inclusion of a prognostic moisture equation. Finally, Diaz and
74 Boos (2019) revisit the potential influence of barotropic instability, and find that in the absence of
75 convective heating, growing disturbances fueled by barotropic instability could be possible with a
76 zonally-uniform basic state; however these disturbances did not grow at rates consistent with ob-
77 served storms, motivating future study in a moist framework. These theories are still young, and
78 their utility for explaining the properties of monsoonal disturbances and their potential sensitivity

79 to changes in the mean state, e.g. induced by increasing greenhouse gas concentrations, has yet to
80 be extensively investigated.

81 The complications of the real world, however, make monsoonal disturbances difficult to study.
82 For instance many comprehensive general circulation models used in the fifth phase of the Cou-
83 pled Model Intercomparison Project (CMIP5) (Taylor et al. 2012) struggle to obtain a realistic
84 distribution of climatological mean June, July, August, and September (JJAS) precipitation rate in
85 the South Asian monsoon region (see the supplement of Sandeep et al. 2018). In addition, sev-
86 eral models run under the Atmospheric Model Intercomparison Project (AMIP) protocol (Gates
87 1992) simulate unrealistic patterns of synoptic activity index (SAI) (also see the supplement of
88 Sandeep et al. 2018), a metric that quantifies an intensity-weighted frequency of MLPS days per
89 season at each location (Ajayamohan et al. 2010). To some extent these errors are attributed to
90 the coarse horizontal resolution of these models; indeed studies have shown that models run with
91 higher resolution such as the UK Met Office’s Unified Model or the Geophysical Fluid Dynam-
92 ics Laboratory’s (GFDL’s) HiRAM demonstrate increased skill in simulating MLPSs (Hunt and
93 Turner 2017; Sandeep et al. 2018).

94 Despite sometimes having errors in the exact location of storms, however, some coarse-
95 resolution General Circulation Models (GCMs) (such as GFDL’s AM4) have been shown to have
96 the ability to reasonably simulate their general frequency statistics and structure (Adames and
97 Ming 2018a), indicating that exact-realism of precipitation location and mean winds is not nec-
98 essarily required for studying the structure and propagation of these dynamical phenomena. It
99 prompts the question of whether a simpler model, lower in the complexity hierarchy, could cap-
100 ture the essence of MLPSs. By a simpler model, we mean one somewhere in between an idealized
101 aquaplanet GCM (like Frierson et al. 2006) and a comprehensive GCM (complete with intricate
102 parameterizations of convection, clouds, radiation, land, chemistry etc.). Xie and Saiki (1999), for

instance, found abundant westward-propagating cyclonic vorticity anomalies, akin to MLPSs, in a simulation using a very coarse horizontal resolution GCM (T21 spectral truncation) and heavily simplified lower boundary conditions. It is worth revisiting these disturbances in a similar setup in light of recent developments (e.g. Boos et al. 2015; Cohen and Boos 2016; Hunt and Parker 2016; Adames and Ming 2018b,a; Diaz and Boos 2019).

In this study we start from a version of Frierson et al. (2006)’s idealized moist model coupled to a full radiative transfer code (Clark et al. 2018), and build up in complexity to attain an environment capable of supporting MLPS-like disturbances. We use this setup, coupled with detailed analysis of the composite anomalous budgets of Ertel potential vorticity, relative vorticity, column internal energy, and column moisture, to discuss the potential applicability of the theories for MLPS propagation described above, and touch on the importance of various boundary conditions, like topography, in the realism of the disturbances simulated.

2. Methods

a. Model description

The modeling setup we use to simulate MLPSs is heavily idealized. Our starting point is the GFDL idealized moist model as configured in Clark et al. (2018). This global model was first introduced in Frierson et al. (2006, 2007), where it consisted of a spectral dynamical core, with simplified moist physics, boundary layer, and radiation parameterizations. It has since been modified to include a simplified Betts-Miller moist convection scheme (Frierson 2007b)¹, alterations to

¹An important parameter in this convection scheme is the relaxation time; we use a relaxation time of 2 h, which is the typical default value used by other studies (e.g. Frierson 2007b; O’Gorman and Schneider 2008; Geen et al. 2017). While it has been shown that aspects of the climate can be sensitive to this parameter choice (e.g. Frierson 2007a; Clark et al. 2018), we find qualitatively similar results in an experiment with a larger convective relaxation time (12 h; not shown), which has been suggested by Bretherton et al. (2004) to potentially be a more appropriate convective relaxation time scale for convection parameterizations of the type used in this model.

122 the boundary layer scheme (O’Gorman and Schneider 2008), and an option to run with full radia-
123 tive transfer, rather than the original gray radiative transfer scheme (Clark et al. 2018). While the
124 full radiative transfer scheme interacts with the active water vapor tracer in the model, there is no
125 parameterization of cloud condensate, and therefore no cloud radiative effects or feedbacks. Slab
126 ocean aquaplanet configurations similar to this, i.e. full radiative transfer with simplified moist
127 physics, have been used before, e.g. in Merlis et al. (2012a,b), Jucker and Gerber (2017), and
128 Vallis et al. (2018).

129 In this study, we examine MLPs in the South Asia region. Due to the annual cycle in solar in-
130 solation, these primarily occur in the boreal summer months of June, July, August, and September,
131 but can also occur during other parts of the year (Hurley and Boos 2015). To capture this seasonal
132 variation in climate, we run all of our simulations with Earth’s current approximate obliquity
133 and eccentricity parameters, 23.439° and 0.01671 respectively. In addition we introduce a crude
134 parameterization of land. In prior studies, land has been added to variants of this model with vary-
135 ing degrees of complexity depending on the application, typically involving modification of some
136 combination of the heat capacity, evaporation parameterization, surface roughness, surface albedo,
137 and surface height over the land portion of the domain (e.g. Byrne and O’Gorman 2012; Merlis
138 et al. 2012b; Maroon et al. 2016; Maroon and Frierson 2016; Voigt et al. 2016; Geen et al. 2017;
139 Vallis et al. 2018; Zhou and Xie 2018). In other models, simplified land has been added in similar
140 ways (e.g. Xie and Saiki 1999; Becker and Stevens 2014; Cronin et al. 2015). As a starting point
141 in our model we choose to distinguish land from the default lower boundary, a slab ocean, in only
142 two ways: its heat capacity, and its treatment of evaporation.

143 The land setup maintains the slab ocean model across the entire lower boundary; however, over
144 land grid cells we use a shallower mixed layer depth, which controls the heat capacity, and scale
145 the potential evaporation rate as predicted by the bulk formula over a saturated surface by a fraction

determined using a simple bucket hydrology model, the same as described in Vallis et al. (2018), which is similar to that in Byrne and O’Gorman (2012) or Zhou and Xie (2018), which dates back to Manabe (1969). The mixed layer depths over land and ocean are the same as those used in experiments in Geen et al. (2017), 2 m over land and 20 m over ocean, and the bucket hydrology model parameters are the same as those described in Vallis et al. (2018), a bucket depth of 150 mm and a bucket saturation fraction of 0.75. We use a surface albedo of 0.26 over land and ocean. The global mean surface albedo is greater than it might be in a comprehensive GCM as to increase the planetary albedo in the absence of clouds (Frierson et al. 2006). Finally, we prescribe zero heat flux from the ocean to the atmosphere, meaning, as a simplification, we assume the ocean does not facilitate any horizontal energy transport. We assume the same for land regions.

In principle one could tune the mixed layer depths, surface albedo, bucket hydrology scheme parameters, and prescribed ocean heat fluxes [e.g. following the procedure outlined in Vallis et al. (2018)] to produce a mean state climate as close as possible to that in observations. However, to maintain a connection to simpler configurations we elect to use the setup described above, which produces a mean state climate that resembles that in observations, but is not an exact match (e.g. it does not contain a significant Western Pacific Warm Pool signature, or east-west asymmetry in ocean basin SST due to warm western boundary currents). For a comparison of the mean state in our simulation to that in observations, see Figure S1, which compares the global pattern in JJAS mean precipitation rate and surface temperature in our idealized simulation to that in Tropical Rainfall Measurement Mission (TRMM) observations (Huffman et al. 2007) and ERA-Interim reanalysis (Dee et al. 2011).

With regard to atmospheric composition, we use approximately present-day concentrations of the well-mixed greenhouse gases ($\text{CO}_2 = 369.4$ ppm, $\text{CH}_4 = 1.821$ ppm), and prescribe a

hemispherically-symmetric pattern of ozone, based on the pattern used in the Aqua-Planet Model Intercomparison Project (Blackburn et al. 2013).

Similar to Geen et al. (2017), to improve the numerical stability of the dynamical core in the upper levels of the model, we add a Rayleigh damping tendency to the horizontal winds. The Rayleigh damping coefficient we use decreases faster than linearly from a value near 0.33 d^{-1} at the top of the model to near zero near the surface, following the vertical profile defined in Equations 13.89 and 13.90 in Jablonowski and Williamson (2011), which were first used in Boville (1986). This Rayleigh damping profile was used for several years in the European Centre for Medium-Range Weather Forecasts (ECMWF) Integrated Forecast System (IFS) model (Jablonowski and Williamson 2011).

b. Experiments

In this study we focus on a simulation with “land” as described in Section 2.a, with realistic continental geometry, but flat topography. We run the model for 20 years, starting from spatially-uniform initial conditions (constant initial temperature and specific humidity), storing 6-hourly mean values of the relevant diagnostics. After the first 10 years, the model approximately reaches equilibrium, here defined as the moment at which the annual global mean net top of atmosphere radiative flux begins to hold steady at near zero. Accordingly, we use the final ten years of each simulation for analysis. The model is configured with 40 unevenly-spaced vertical sigma levels, with approximately three levels within the planetary boundary layer, and extends to the top of the atmosphere, with a top-level interface pressure of 0 hPa. In the horizontal, we run the model at T42 spectral resolution, which corresponds to approximately $2.8^\circ \times 2.8^\circ$ horizontal resolution in grid space.

191 As part of this study, we ran two other simulations flanking the simulation described above in
192 terms of the complexity of the lower boundary, though we do not show their results. On the simpler
193 side, we ran an aquaplanet case with a uniform slab ocean mixed layer depth of 20 m, while on the
194 more complex side, we ran a case with land and realistic topography, spectrally regularized as in
195 Lindberg and Broccoli (1996). The aquaplanet case produced an annual cycle in precipitation sig-
196 nificantly lagged from that on Earth's, with monthly mean precipitation rates maximizing during
197 September and October at the latitudes of the South Asian monsoon region. It also lacked a real-
198 istic polewardly-increasing meridional temperature gradient and attendant easterly vertical wind
199 shear². On the other hand, the case with realistic topography produced MLPSs, but despite regu-
200 larization, suffered from severe spectral ringing in the mean precipitation field near the Himalayas,
201 compromising the quality of the climate relative to the flat topography case. Therefore the config-
202 uration with land and flat topography happened to be the best configuration of this model we tested
203 for studying South Asian MLPSs. It is possible that one could obtain MLPS-like disturbances in a
204 simulation with further idealized continental geometry and surface hydrology. However we leave
205 such experimentation to future work and choose to focus on our realistic continental geometry,
206 flat topography simulation, which provides an important link between MLPSs in comprehensive
207 GCMs and those in more idealized frameworks.

²One could potentially produce a more realistic meridional temperature gradient, and potentially MLPSs, in the aquaplanet configuration of this model by removing the seasonal cycle in solar insolation and adding a forcing to induce a warm pool at the latitudes of the South Asian monsoon region, e.g. centered at around 15°N. For instance Ajayamohan et al. (2014) find westward propagating Rossby-wave-like disturbances in such a setup; however, in this study we opt to allow the natural interaction between the seasonal cycle and land-ocean contrast in heat capacity to produce such a local temperature maximum in the region.

208 *c. Analysis techniques*

209 To analyze the structure of MLPSs in our model, we employ frequency-wavenumber spectral
210 analysis, compute lag regression patterns, and tracer budgets. Frequency-wavenumber spectral
211 analysis allows us to identify the frequencies and wavenumbers of the zonally-propagating waves
212 that are most prevalent; this type of analysis is commonly used in studying equatorial waves (e.g.
213 Wheeler and Kiladis 1999; Hendon and Wheeler 2008), though here we apply the technique in
214 the latitudes of the South Asian monsoon region. Lag regression patterns allow us to determine
215 the spatial structure of variable anomalies projected onto a MLPS index. Tracer budget analysis
216 allows us to determine the leading terms governing the evolution of MLPSs in our model. We
217 approximately follow the methods described in Adames and Ming (2018a). Here we will explain
218 the details of these techniques which we will employ later.

219 1) SPECTRAL ANALYSIS

220 To compute frequency-wavenumber power spectra, we start with 6-hourly resolution model out-
221 put of the precipitation rate. We then subset this dataset in time such that it only includes datapoints
222 for the months of June, July, August, and September. From this timeseries, we construct a set of
223 60-day segments, which overlap by 30 days, generating a four-dimensional dataset, with dimen-
224 sions time, longitude, segment, and latitude; it follows that the segment dimension has length 242.
225 We apply a Hanning window over the time dimension, tapering the endpoints of the segments
226 toward zero to minimize spectral leakage (Welch 1967); in addition, we apply a Hanning win-
227 dow over 50°E to 130°E to taper data to zero outside our longitudinal region of interest. After
228 this preparation, we compute a fast Fourier transform (FFT) in longitude and time, and compute
229 the power as the square of the magnitude of the complex Fourier coefficients. To construct a
230 two-dimensional frequency-wavenumber diagram, we average the power over the segments and

231 between latitudes bounding the region of interest for the particular dataset, which correspond
 232 roughly to the latitudinal bounds of the South Asian monsoon region, and then compare it to a
 233 reference red frequency spectrum. We define the region of interest for a particular dataset as $\pm 5^\circ$
 234 from the latitude of maximum mean JJAS precipitation rate³ along the 80°E longitude band. We
 235 compute the red spectrum as in Masunaga et al. (2006), normalizing such that the sum of the power
 236 in non-zero frequencies matches that in the power spectrum of the precipitation rate.

237 To compare the power in the signal to that in the reference red spectrum, we compute what is
 238 referred to as the “signal strength” (S) by determining the ratio of the difference between the power
 239 spectrum (P) and red spectrum (R) to the power spectrum itself:

$$S = \frac{P - R}{P} = \frac{P/R - 1}{P/R}. \quad (1)$$

240 Statistical significance is determined by computing a critical value of a chi-squared-statistic at the
 241 99% significance level, which is the ratio of two variances scaled by the degrees of freedom (n)
 242 minus one, e.g. $\chi^2 = \frac{P(n-1)}{R}$. The number of degrees of freedom used in computing the critical
 243 chi-squared value is calculated as in Hendon and Wheeler (2008) and Adames and Ming (2018b);
 244 it is equal to 2 (amplitude and phase) $\times$ 10 (number of years) $\times$ 122 (number of days in JJAS per
 245 year) / 60 (days per segment) $\approx$ 40. At the 99% level, this results in a critical χ^2 -value of 62.4,
 246 indicating that if the power of the signal is 1.6 times that of the red spectrum then there is a 1%
 247 chance the signal emerged out of red noise. In terms of the signal strength in Equation 1, this
 248 means in order for the signal to be statistically significant at the 99% level, the signal strength
 249 must be greater than or equal to approximately 0.38:

$$\frac{\chi^2_{\text{critical}}}{n - 1} = \left(\frac{P}{R} \right)_{\text{critical}} = 1.6 \implies S_{\text{critical}} = 0.38. \quad (2)$$

³Hurley and Boos (2015) note that MLPS activity is strongest slightly poleward of this maximum in most monsoon regions; however we claim that as a first approximation this is a reasonable method of defining the central latitude of our region of interest.

250 2) LAG REGRESSION ANALYSIS

251 To compute lag regression patterns we follow the methods of Adames and Wallace (2014) and
252 Adames and Ming (2018a). This requires computing an index, which measures the intensity of
253 MLPS activity. Adames and Ming (2018a) do this by spectrally filtering the precipitation rate to
254 include wave activity from only MLPS-like modes (zonal wavenumbers, k , between -25 and -3
255 and frequencies greater than 0.067 d^{-1}), then averaging over the spatial region of interest, here
256 defined as $\pm 5^\circ$ latitude from the latitude of maximum JJAS mean precipitation rate, between
257 75°E and 85°E in longitude; this results in a one-dimensional index over time, which is then
258 standardized such that it has a mean of zero and a standard deviation of one, yielding a vector $\mathbf{P}^T$.
259 Here k is the non-dimensional zonal wavenumber, which can be related to a dimensional zonal
260 wavenumber $\tilde{k}$ via $k = \tilde{k}a \cos \phi$, where a is the radius of the Earth and ϕ is latitude. The spectral
261 filtering is achieved by performing standard Fourier transforms in time and longitude of the raw
262 precipitation rate timeseries, zeroing out all coefficients outside of the rectangular spectral region
263 specified above, and computing an inverse Fourier transform back to time and longitude space.

264 With an index in hand, we can then regress any variable against it. Borrowing notation from
265 Adames and Wallace (2014) this looks like:

$$\mathbf{D} = \frac{\mathbf{S}\mathbf{P}^T}{N}. \quad (3)$$

266 Here $\mathbf{S}$ is a two-dimensional matrix with each row representing the time series of a variable at
267 a given gridcell. To ensure that we are capturing the anomalies associated with high-frequency,
268 i.e. storm-timescale, variability, we spectrally filter all quantities we regress such that they contain
269 frequencies of only 0.067 d^{-1} and above; this approach is analogous to the approach used in Kim
270 et al. (2013), who filtered the meridional advection term of the column-integrated MSE budget
271 to periods between 20 d to 100 d before regressing, to isolate MJO-related anomalies. $\mathbf{P}$ is the

272 standardized index at each time, i.e. it is a single row vector; N is the number of values in the
273 index; and $\mathbf{D}$ is the computed regression pattern. $\mathbf{D}$ contains a time-independent spatial pattern of
274 anomalies with the same spatial dimensions as the input variable. Lag regressions can be computed
275 by shifting the index forward or backward in time and applying the same procedure, noting that
276 this reduces the number of overlapping elements between the index and variable, i.e. it slightly
277 changes N . This allows us to construct a picture of what the conditions look like before, during,
278 and after a monsoon low pressure system event occurs.

279 To smooth out regression patterns, particularly in the context of the tracer budgets, we apply a
280 regression-compositing technique similar to the one employed in Adames and Ming (2018a). This
281 entails computing regression patterns for index regions shifted -2, -1, 0, 1, or 2 grid cells away
282 in longitude and/or latitude from the original center of the region of interest described above, and
283 then shifting the regression patterns back to all be centered at the same location and averaging.
284 This results in computing and taking the mean of 25 regression patterns, producing a smoother
285 picture.

286 3) TRACER BUDGET ANALYSIS

287 We compute budget terms for four equations in this study: Ertel potential vorticity (EPV), rel-
288 ative vorticity, column-integrated internal energy, and column-integrated moisture, the equations
289 for which are given and discussed in Section 3. Anomalous terms for each budget are computed
290 by regressing each time series against the precipitation index defined above. In the case of EPV
291 and relative vorticity, four-dimensional fields, the time series of each term is computed explicitly;
292 where needed, second-order finite differences are used to estimate partial derivatives in the inte-
293 rior, and first-order finite differences are used to estimate partial derivatives on the boundaries. For

294 computing horizontal derivatives on the sphere, we follow the methods of Seager and Henderson
295 (2013). For EPV and relative vorticity, the residuals from computing the terms explicitly are small.

296 For the column-integrated fields of the internal energy and moisture budgets, using a purely
297 explicit procedure does not result in an adequately closed budget. In those cases, as an objective
298 way of partitioning the residual, we follow the methods of Hill et al. (2017)⁴. This entails first
299 computing an adjusted set of horizontal winds at each vertical level using the flux-form framing
300 of the column-integrated budget to ensure things are balanced. We then use these adjusted winds
301 to compute the horizontal advection terms in the advective form of the budget, and compute the
302 column-integrated vertical advection term as a residual.

303 A useful tool to quantify the extent to which a term in an anomalous budget (X') contributes to
304 the anomalous time tendency of a tracer (m) is projection analysis. This is a technique that has
305 been used frequently in prior studies (e.g. in Andersen and Kuang 2011; Lutsko 2017; Adames
306 and Ming 2018a). It entails computing the integral of the product of the term X' with the time
307 tendency anomaly term $\frac{\partial m'}{\partial t}$ over a region A , then dividing by the integral of the square of the
308 tendency over the same region:

$$\mathcal{P}_{X'} = \iint_A X' \frac{\partial m'}{\partial t} dA \Big/ \iint_A \left(\frac{\partial m'}{\partial t} \right)^2 dA. \quad (4)$$

309 Here as A we use the rectangular region 50°E to 110°E, 0° to 30°N, to approximately enclose the
310 South Asia horizontal region. Since we are integrating only over the horizontal dimensions, if the
311 terms are defined in the vertical, the projection, $\mathcal{P}$ will be a function of pressure, e.g. for EPV;
312 otherwise the projection will be a scalar quantity, e.g. for column-integrated internal energy.

⁴Note in the case of the internal energy budget the methods of Hill et al. (2017) cannot be applied exactly; therefore the budget can still not be exactly closed. Instead we follow the method described in the Appendix.

3. Results

a. Mean state climate

The observed JJAS mean state climate in the South Asian monsoon region has a number of distinctive attributes (Sikka 1977). We illustrate these attributes and their counterparts in the simulation in the two columns of Figure 1. Given the heavily-idealized nature of the simulation we do not expect an exact match to the real-world climate; however, we find in our simulation that several theoretically-important broad-scale features are obtained.

We first look at the JJAS-mean precipitation rate, panels (a) and (b) in Figure 1. Figure 1(a) corresponds to the 2001-2016 JJAS-mean precipitation rate computed using observations from the Tropical Rainfall Measurement Mission (TRMM) (Huffman et al. 2007) and panel (b) corresponds with JJAS-mean precipitation from our simulation. In observations, while the detailed structure is strongly influenced by small-scale features of the topography, the large-scale structure roughly corresponds in a local shift of the Intertropical Convergence Zone (ITCZ) northward to around 20°N. This northward shift in the area of maximum time mean precipitation is roughly captured in the simulation, albeit to a lesser extent. Figure 1(b) shows that JJAS-mean precipitation maximizes at roughly 12.6°N in our simulation, with a secondary maximum located near the equator⁵.

Another notable feature of the observed climate is that column-integrated moisture increases steadily as one moves northward from the equator through the Bay of Bengal (Adames and Ming

⁵A similar double Intertropical Convergence Zone (ITCZ) structure was also found in Xie and Saiki (1999). In their idealized simulation they also found latitudinal maxima in summer precipitation in the South Asian monsoon region and near the equator. To a lesser extent a double ITCZ structure is also seen after monsoon onset in the idealized “flat” simulation of Geen et al. (2017), which uses a similarly-configured model to our simulation, with the single exception of additional of AMIP-derived slab ocean heat fluxes in their case. In our case, the double ITCZ structure is not a completely persistent feature. In other words, there are times when it is only raining in one rainband and not the other, but during July and August there are times when it rains in both rainbands simultaneously.

2018a). This is illustrated by plotting 2001-2016 JJAS-mean column-integrated water vapor from the ERA-Interim reanalysis dataset (Dee et al. 2011) in Figure 1(c). This strong gradient has been theorized to play a role in the dynamics of monsoon low pressure systems (Adames and Ming 2018b). Figure 1(d) shows the JJAS mean column-integrated moisture in our simulation. There we can see a band of high column-integrated water vapor roughly coincident with the band of high precipitation rate, running from the Arabian Sea, across India, and over the northern Bay of Bengal and Southeast Asia. Compared with reanalysis, where the water-vapor maximum is located near the land-sea boundary between Bangladesh and the Bay of Bengal, the water-vapor maximum in our simulation is displaced slightly southward. In addition, column-integrated moisture magnitudes are substantially smaller than those seen in reanalysis, with maximum values of around 20 mm in our idealized simulation and around 60 mm in reanalysis (Adames and Ming 2018a).

The third salient property of the mean state South Asian monsoon climate is a meridionally-increasing surface temperature field, and attendant easterly vertical wind shear, with westerly winds near the surface and easterly winds aloft (Xie and Saiki 1999; Boos et al. 2015; Cohen and Boos 2016). This is illustrated using 2001-2016 JJAS-mean 600 hPa temperature, and the difference between the 200 hPa and 850 hPa winds from the ERA-Interim reanalysis dataset (Dee et al. 2011) in Figure 1(e). The meridionally-increasing temperature gradient is induced by the difference in heat capacity between the land and ocean. Because the land heats up faster than the ocean, it experiences greater seasonal variation in surface temperatures than ocean at similar latitudes, which gets communicated to the free troposphere. Through thermal wind balance, this positive meridional temperature gradient is associated with easterly vertical wind shear (Vallis 2006). The crude setup in our simulation is able to capture this, as indicated by the polewardly-increasing 600 hPa temperature in Figure 1(f) and quiver arrows pointing from the east to the west, indicative of easterly vertical wind shear, with strongest values of about 30 m s^{-1} at around

355 12.5°N. The magnitude of the shear decreases as one moves northward over the Asian continent,
356 which is similar to what is seen in reanalysis, e.g. in Figure 1(e).

357 *b. The general character of South Asian monsoon low pressure systems in the idealized simulation*

358 1) DOMINANT ZONAL WAVENUMBER AND FREQUENCY OF PRECIPITATING DISTURBANCES

359 Despite the simplicity of the setup of the idealized simulation, notably omitting the impacts of
360 the prominent land surface topography of Southern Asia, and the impacts of ocean heat trans-
361 port, we seem to obtain an adequate South Asian JJAS mean state climate to support westward-
362 propagating, precipitating disturbances. This can be made immediately apparent by looking at
363 a time-longitude diagram of unfiltered precipitation during an example summer season averaged
364 between 7.6°N and 17.6°N, the colors in Figure 2. To guide one's eye, contour lines are added,
365 which indicate the value of the precipitation rate filtered to include data only from MLPS-like
366 modes, i.e. with zonal wavenumbers between -25 and -3 and frequencies greater than 0.067 d^{-1} ,
367 using the method described in Section 2.c2. In the particular season shown in Figure 2, we find
368 active westward-propagating MLPS-like activity during July, a break in August, and reinvigorated
369 activity in September.

370 To determine the characteristic frequency and zonal wavenumber of the disturbances, we can
371 compute a frequency-wavenumber power spectrum of the precipitation rate. For the idealized
372 simulation this is shown in Figure 3(b). There we find statistically significant signal strength be-
373 tween zonal wavenumbers -20 to -5 , and frequencies 0.10 d^{-1} to 0.35 d^{-1} . This pattern in signal
374 strength is largely consistent with that seen in daily precipitation rate observations from the Tropi-
375 cal Rainfall Measurement Mission (TRMM) (Huffman et al. 2007) [Figure 3(a)] and a simulation

376 using GFDL's AM4 (Adames and Ming 2018a) [cf. Figure 3 of Adames and Ming (2018a)]⁶, and
 377 is indicative of westward-propagating waves of alternating wet and dry periods with a horizon-
 378 tal scale on the order of 1000 km and a period of around 3 d to 10 d. A characteristic frequency,
 379 $\overline{f_w}$, and zonal wavenumber, $\overline{k}$, can be computed by taking statistically-significant-signal-weighted
 380 means of each over the plotted domain in Figure 3:

$$\overline{f_w} = \frac{\int \int_{S \geq S_c} S f_w df dk}{\int \int_{S \geq S_c} S df dk} \quad (5)$$

$$\overline{k} = \frac{\int \int_{S \geq S_c} S k df dk}{\int \int_{S \geq S_c} S df dk}, \quad (6)$$

382 where S is the signal strength, and S_c is the critical signal strength for statistical significance. In
 383 the observations this results in $\overline{k} = -8.7$ and $\overline{f_w} = 0.20 \text{ d}^{-1}$, while in the idealized simulation this
 384 results in $\overline{k} = -10.5$, $\overline{f_w} = 0.17 \text{ d}^{-1}$.

385 2) HORIZONTAL STRUCTURE OF PRECIPITATION, MID-LEVEL VERTICAL VELOCITY, AND 386 LOW-LEVEL WIND ANOMALIES

387 The structure of the westward-propagating disturbances can be elucidated using regression anal-
 388 ysis as described in Section 2c, following the methods of Adames and Ming (2018a). We will
 389 first consider the horizontal structure of the anomalous precipitation, mid-level (500 hPa) verti-
 390 cal velocity, low-level (850 hPa) wind fields on days preceding, during, and following a storm
 391 event centered at 80°E and the latitude of maximum mean JJAS precipitation along 80°E in the
 392 South Asian monsoon region. In the idealized simulation, as we look at the lag sequence de-
 393 scending from the top of Figure 4, we can see clear evidence of a westward-propagating cyclonic
 394 disturbance crossing the Bay of Bengal and traversing India over a span of about four days. The
 395 disturbance is flanked by dry anticyclonic circulations. Vertical velocity anomalies slightly lead

⁶Note that the region of interest used here and in Adames and Ming (2018a) for the observations was centered at 17.5°N, rather than at 12.6°N
 in the case of the idealized simulation in Figure 3(b).

precipitation anomalies, similar to the disturbances analyzed in Adames and Ming (2018b). The maximum low-level wind speed anomaly associated with the regression in the idealized experiment at lag day zero is 2.2 ms^{-1} . This value is weaker than the observed wind perturbation, which is greater than 8.5 ms^{-1} (Hurley and Boos 2015). This is due in part to the implicit compositing effect of the regression analysis. Individual storms of a similar strength to the observations do occur in the idealized simulation⁷; however a more careful analysis involving tracking individual storms would be required to comprehensively compare the intensity distribution of the storms in our idealized simulation to that in observations.

In comparison to regression results from GFDL's AM4, the disturbances are located farther south and have weaker precipitation anomalies, on the order of 4 mm d^{-1} versus 10 mm d^{-1} , but have similar magnitude wind anomalies. The propagation direction is almost directly westward, the same direction as the climatological vertical wind shear [Figure 1(h)], raising the possibility that the disturbances could be adiabatically steered by the climatological mid-tropospheric winds (Boos et al. 2015).

The propagation velocity of the storms can be quantified by computing the location of the maximum vorticity anomalies at 850 hPa at each lag day. This is done by interpolating the vorticity anomalies from the model native grid, spaced by roughly 2.8° in latitude and longitude, to a $0.1^\circ \times 0.1^\circ$ grid, using the nonlinear “patch” interpolation method provided by the Earth System Modeling Framework (ESMF) (Collins et al. 2005; Zhuang and Jüling 2019). The maximum at each lag day plotted in Figure 4 is marked with a filled black circle. We can compute an average zonal and meridional propagation velocity over the four-day window plotted in Figure 4 for each

⁷For instance, if we compute composite means of the anomaly patterns associated with precipitation index values greater than two (approximately the strongest 3–4% of storms), we find storm-center precipitation anomalies on the order of 10 mm d^{-1} , maximum wind speed anomalies near 8.5 ms^{-1} , and minimum surface pressure anomalies of less than 3.6 hPa. The results of this composite analysis are shown in Figure S2 of the supplement.

simulation by taking the difference in the position of the maximum at lag day 2 and the position of the maximum at lag day -2 and dividing by the difference in time (4 d). If we do this, we find that the average zonal propagation velocity of the maximum in the idealized simulation is -5.9 ms^{-1} , while the average meridional propagation velocity is -0.1 ms^{-1} . The propagation velocity in our simulation is stronger and more westward-directed than in reality (Boos et al. 2015) or in comprehensive GCMs (Adames and Ming 2018a); in both, zonal propagation velocities are typically on the order of 4 ms^{-1} or smaller, and at least in Boos et al. (2015) there is a more significant meridional component.

c. The dynamical properties of South Asian monsoon low pressure systems in the idealized simulation

1) VERTICAL STRUCTURE OF ERTEL POTENTIAL VORTICITY ANOMALIES

Recent studies have illuminated the composite vertical structure of Ertel potential vorticity (EPV) anomalies of observed MLPSs in both the pressure-longitude and pressure-latitude planes (e.g. Boos et al. 2015; Cohen and Boos 2016). The structure of EPV can inform us about a disturbance's dynamics, e.g. Boos and Korty (2016) look at a zonal profile to assess whether moist baroclinic instability is occurring in observed MLPSs and Boos et al. (2015) look at a meridional profile to assess whether observed MLPSs could be steered westward by the JJAS mean zonal winds. To begin understanding the dynamics of the MLPSs in our idealized simulation, we plot such profiles as well.

We compute EPV on isobars following Tamarin and Kaspi (2016):

$$q_d = -g \left(f \hat{\mathbf{k}} + \nabla \times \mathbf{u} \right) \cdot \nabla \theta, \quad (7)$$

where all horizontal derivatives are computed on surfaces of constant pressure, and in spherical coordinates. $\mathbf{u} = (u, v, \omega)$ is the three-dimensional wind velocity in pressure coordinates, $\hat{\mathbf{k}}$ is the vertical unit vector, f is the Coriolis parameter, g is the gravitational acceleration, and θ is the potential temperature. Horizontal derivatives are computed using second-order centered finite differences following the methods described in Seager and Henderson (2013). Vertical derivatives are computed using second-order centered finite differences in the interior and first order finite differences on the boundaries. We scale q_d by 10^6 such that it has units of potential vorticity units (PVU).

If we compute EPV using six-hourly output, regress it onto the precipitation index at lag day zero, and average the result over the latitudes of the region of interest, the result is Figure 5(a), a zonal cross-section of anomalous EPV for MLPSs in our idealized simulation. Overlaid are contours representing a similarly obtained cross section of temperature anomalies. Associated with MLPSs we find a slightly westward-tilting, i.e., in the direction of the shear vector plotted in Figure 1(f), column of anomalous positive EPV centered at around 80°E and 600 hPa. The positive EPV anomalies are flanked to the west and east by weaker negative EPV anomalies, with similar tilts. Positive temperature anomalies can be found above and slightly to the east of the positive EPV anomalies, centered at around 82°E and 400 hPa, while negative temperature anomalies can be found to the west and east as well as below the central warm core. While the anomalies are weaker, the zonal profiles of temperature and EPV anomalies obtained in our idealized simulation qualitatively match those of observed MLPS shown in Cohen and Boos (2016), though they lack the characteristic bimodality of the EPV profile of MLPSs in the vertical noted by Hurley and Boos (2015) and Cohen and Boos (2016) in the ERA-Interim reanalysis dataset and Hunt and Turner (2017) in simulations using the Met Office Unified Model (MetUM). This could be due in part to the coarser vertical resolution used in our idealized simulation (40 vertical levels), when compared

with the vertical resolution used in ERA-Interim (60 vertical levels) (Dee et al. 2011) or MetUM (85 vertical levels) (Hunt and Turner 2017).

Taking the same EPV anomalies obtained through regression, but this time averaging between 75°E and 85°E, we can obtain a profile of the composite disturbance in the pressure-latitude plane. This is shown in Figure 5(b). Following Boos et al. (2015) we also plot contour lines representing the mean JJAS zonal winds averaged over the same longitudinal region as the EPV anomalies⁸. In Figure 5 we find that the bulk of the disturbance, identified as the region of positive EPV anomalies, is located in a region of westerly winds (solid contours). At upper levels, there is some overlap between positive EPV anomalies and easterly winds; however, the magnitude of the easterly winds are fairly weak in the region of overlap.

2) ANOMALOUS ERTEL POTENTIAL VORTICITY BUDGET

We can learn more about the propagation mechanism of the MLPSs in our idealized simulation by computing anomalous tracer budgets. We will start with the anomalous EPV budget associated with the storms; a similar analysis was done for a single particular storm in Boos et al. (2015), though here our anomalies represent a composite obtained through regression.

An equation governing the time tendency of EPV is given in Tamarin and Kaspi (2016):

$$\left(\frac{\partial q_d}{\partial t}\right)' = -(\mathbf{u} \cdot \nabla q_d)' - g \left[\boldsymbol{\eta} \cdot \nabla \left(\frac{d\theta}{dt} \right) \right]' . \quad (8)$$

The first term on the right hand side of Equation 8 corresponds with the EPV tendency associated with advection of EPV and the second term corresponds to the EPV tendency due to diabatic processes. $\boldsymbol{\eta} = f\hat{\mathbf{k}} + \nabla \times \mathbf{u}$ is the three-dimensional absolute vorticity vector. The advection term

⁸Note that in Boos et al. (2015) the zonal wind was smoothed with a 21-day running mean and then composited over storms; here we use the full climatological mean rather than a smoothed composite.

480 can be separated into horizontal and vertical components:

$$-(\mathbf{u} \cdot \nabla q_d)' = -(\mathbf{v} \cdot \nabla_p q_d)' - \left(\omega \frac{\partial q_d}{\partial p} \right)' \quad (9)$$

481 It follows that under adiabatic processes, EPV is conserved following the flow.

482 Spatial patterns of the different terms in Equation 8 at the 500 hPa and 700 hPa levels are shown
483 in Figure 6. There we find that the pattern of anomalous EPV time tendency [panels (a) and
484 (f)] is consistent with the westward-propagation of the storms, with positive EPV tendencies to
485 the west of the vortex center and negative EPV tendencies to the east at either level. As Boos
486 et al. (2015) found in a case study of a monsoon depression, in the mid-troposphere a negative
487 diabatic tendency at the storm center [Figure 6(b)] is largely compensated for by a positive vertical
488 advection tendency in the same location [Figure 6(d)]. At this level in Boos et al. (2015) and in
489 our simulation, anomalous horizontal advection of EPV [Figure 6(c)] appears to project most
490 strongly onto to the spatial pattern of the overall EPV tendency. Closer to the surface, at 700 hPa,
491 diabatic processes appear to play a larger role in the propagation tendency [cf. Figure 6(g) and
492 Figure 6(h)], with horizontal advection no longer being as significant; again this is similar to what
493 is found in Boos et al. (2015) in reanalysis. The residual in the budget at each level, shown in
494 Figure 6(e) and 6(j), is small compared to the explicitly computed terms.

495 While the results plotted in Figure 6 provide qualitative evidence of the importance of hori-
496 zontal advection and diabatic processes in the propagation of EPV anomalies, we can be more
497 quantitative about this assessment by applying projection analysis, described in Section 2c, to the
498 EPV budget. The projections for each term at each vertical level in the idealized simulation are
499 shown in Figure 7. Here it is quantitatively clear that anomalous horizontal advection of EPV
500 is dominant in the mid-to-upper troposphere, while diabatic processes, primarily condensation of

501 water vapor⁹, become more important in the lower troposphere, i.e. near 700 hPa. This is qual-
502 itatively consistent with the results of Boos et al. (2015). Vertical advection anomalies have a
503 small negative contribution to the EPV tendency in the lower troposphere and a small positive
504 contribution in the mid-to-upper troposphere; in general they tend to oppose the diabatic tendency
505 throughout the atmosphere. In doing a Reynolds decomposition, a technique described in the next
506 paragraph, of the vertical advection term (Figure S4), we find that at 500 hPa the product of the
507 high frequency vertical velocity and low frequency vertical gradient of EPV term, $\left(\omega' \frac{\partial \overline{q_d}}{\partial p}\right)'$, is the
508 largest term, with the term representing the product between the low-frequency vertical velocity
509 and high-frequency vertical gradient of EPV, $\left(\overline{\omega} \frac{\partial q'_d}{\partial p}\right)'$, being 1/3 of its amplitude, and the other
510 terms playing a negligible roles. At the 700 hPa level, $\left(\overline{\omega} \frac{\partial q'_d}{\partial p}\right)'$ and $\left(\omega' \frac{\partial \overline{q_d}}{\partial p}\right)'$ have a more equal
511 role, with the other terms playing secondary roles. Because, particularly in the mid-troposphere,
512 the spatial pattern of the vertical advection term is largely collocated with the storm-center EPV
513 anomalies, this suggests that the low-frequency vertical gradient in EPV has a role in the growth
514 and maintenance of the storms.

515 Horizontal advection consists of two quadratic terms in the budget, one zonal and one merid-
516 ional. It is worth asking if these terms could potentially be treated as being linear in high-frequency
517 factors, i.e. either linear in a high frequency wind factor or linear in a high-frequency EPV-gradient
518 factor, or whether the anomalous horizontal advection tendency is nonlinear process, i.e. repre-
519 senting advection of high-frequency EPV anomalies by the high-frequency horizontal flow. At
520 least for stronger storms, Boos et al. (2015) suggest that nonlinear processes are at work. To see if

⁹We conclude this by computing the component of the anomalous diabatic EPV tendency due to condensation via Equation 5 of Tamarin and Kaspi (2016), which is based on the work of Emanuel et al. (1987) (Figure S3). We find that it closely resembles the anomalous pattern of the total diabatic EPV tendency over the regression index region, representing 71% of it at the 700 hPa level and 97% of it at the 500 hPa level. The dominance of condensation of water vapor in the diabatic heating component of the EPV tendency is qualitatively consistent with composites of observed MLPSs (Hunt et al. 2016b).

this is the case in our idealized simulation we can perform a Reynolds decomposition on the terms associated with horizontal advection:

$$\begin{aligned}
 -(\mathbf{v} \cdot \nabla q_d)' = & -\frac{1}{a \cos \phi} \left(\bar{u} \frac{\partial \bar{q}_d}{\partial \lambda} + \bar{u} \frac{\partial q_d'}{\partial \lambda} + u' \frac{\partial \bar{q}_d}{\partial \lambda} + u' \frac{\partial q_d'}{\partial \lambda} \right)' \\
 & - \frac{1}{a} \left(\bar{v} \frac{\partial \bar{q}_d}{\partial \phi} + \bar{v} \frac{\partial q_d'}{\partial \phi} + v' \frac{\partial \bar{q}_d}{\partial \phi} + v' \frac{\partial q_d'}{\partial \phi} \right)'. \quad (10)
 \end{aligned}$$

Above we have taken the quadratic advection terms and broken them down into terms that are linear in high-frequency factors and terms that are nonlinear in high-frequency factors. Similar to the method used in (Adames and Ming 2018b), overbars represent low-frequency factors, obtained by spectrally filtering the data to contain frequencies less than 0.067 d^{-1} per day, while primed factors represent the quickly varying residual. Because the overbar factors still vary in time, it is possible the product of two low-frequency fields could project non-negligibly onto the precipitation index¹⁰; therefore we retain those terms in the equation above. Note, as described in Section 2, we filter all products to contain frequencies greater than 0.067 d^{-1} per day before regressing.

Projecting each term onto the total EPV tendency, we find, as was qualitatively shown in Figure 5(b), that advection of high-frequency EPV by the low-frequency zonal wind tends to work against the prevailing westward-propagating tendency of EPV in the lower-to-mid troposphere, the solid red line in Figure 8. Instead, the most important term is the advection of the low-frequency EPV by the high-frequency meridional winds, the dashed blue line in Figure 8. Because the high-frequency meridional winds are cyclonic, they blow southward to the west of the storm, down the low-frequency EPV gradient, bringing high mean EPV air from the north, and northward east of

¹⁰Consider the idealized example of two factors that can be described as a sine and cosine wave of the same frequency, ω . Their product, via the sine double angle identity, would then be a sine wave with twice the frequency, e.g. $\sin(\omega t) \cos(\omega t) = \frac{1}{2} \sin(2\omega t)$, which could project onto a higher frequency index.

the storm up the low frequency EPV gradient, bringing low mean EPV air from the south, resulting in the dipole pattern seen in Figure 6(a). This is consistent with the propagation mechanism of linear Rossby waves.

The other terms, related to advection of low-frequency EPV by the low-frequency horizontal winds, advection of high-frequency EPV by the low-frequency meridional winds, advection of the low-frequency EPV by the high-frequency zonal winds, and advection of the high-frequency EPV by the high-frequency horizontal winds play a negligible role in the propagation of EPV anomalies in the lower troposphere, and play minor offsetting roles in the upper troposphere. The tertiary role of high-frequency EPV advection by the high-frequency horizontal winds in the budget in particular suggests that beta drift is not a primary driver of propagation for the storms in our simulation.

Further insight into the dynamics of the MLPSs in our idealized simulation can be gleaned from analysis of the budgets of vorticity, column-integrated internal energy, and column-integrated moisture. By looking at these budgets, we can approximately determine whether a simplified model, like moisture vortex instability (Adames and Ming 2018a), could be used to describe the storm systems in our simulation.

3) ANOMALOUS VORTICITY BUDGET

The anomalous flux-form vorticity equation discussed in Boos et al. (2015) is given by:

$$\frac{\partial \zeta'}{\partial t} = - \left[\nabla \cdot (f + \zeta) \mathbf{u} \right]' - \nabla \cdot \left(\omega \hat{\mathbf{k}} \times \frac{\partial \mathbf{u}}{\partial p} \right)'. \quad (11)$$

In this budget, $-\nabla \cdot f \mathbf{u}$ represents the collective influence of vortex stretching and horizontal advection involving the planetary vorticity, $-\nabla \cdot \zeta \mathbf{u}$ represents the collective influence of vortex stretching and horizontal advection involving the relative vorticity, and $\nabla \cdot \left(\omega \hat{\mathbf{k}} \times \frac{\partial \mathbf{u}}{\partial p} \right)$ represents the collective influence of vertical vorticity advection and vortex tilting. It is useful to view the

spatial anomaly patterns of the budget terms through this decomposition, because it separates terms into two categories. The first category consists of terms that are definitively linear in high-frequency factors and are not impacted by the low-frequency winds; only $-\nabla \cdot f\mathbf{u}$ belongs in this category¹¹. The second category consists of terms that may contain components nonlinear in high-frequency factors and/or influence of the low-frequency winds; $-\nabla \cdot \zeta\mathbf{u}$ and $\nabla \cdot \left(\omega\hat{\mathbf{k}} \times \frac{\partial\mathbf{u}}{\partial p}\right)$ are the terms in this category.

The terms in the anomalous budget for a level in the upper troposphere (400 hPa) and a level in the lower troposphere (850 hPa), decomposed as described above, are shown in Figure 9. Panels (a) and (e) shows the anomalous time tendency of the relative vorticity. There we can see a dipole pattern oriented along an east-west axis, similar to what we see in the anomalous EPV budget. In addition we can see that the dominant term on the right hand side of Equation 11 is the term involving the planetary vorticity. Terms potentially involving the low-frequency winds, i.e. $-\nabla \cdot (\zeta\mathbf{u})'$ or $-\nabla \cdot \left(\omega\hat{\mathbf{k}} \times \frac{\partial\mathbf{u}}{\partial p}\right)'$, are about an order of magnitude smaller at both 850 hPa and 400 hPa, and to some extent offset each other.

Figure 10 shows the projection of the vorticity budget terms on the time tendency of relative vorticity in the idealized simulation. There we can see that our budget closes nearly perfectly below about 500 hPa and only slightly diverges above, as evidenced by the dashed black line, representing the total of the terms on the right hand side of Equation 9 having a projection of about one at all pressure levels. In addition, we see quantitative evidence of the dominance of the planetary vorticity term, the red line in Figure 10, which indicates a spatial projection of over 0.5 below 300 hPa. It is only above 300 hPa anomalous vortex stretching associated with the relative vorticity and/or anomalous relative vorticity advection, $-\nabla \cdot \zeta\mathbf{u}$, becomes of leading-

¹¹Note that since we highpass filter all quantities before regressing, as described in Section 2.c, any low-frequency component of $\mathbf{u}$ will be filtered out.

order significance in the budget. The combined effects of anomalous vertical advection and vortex tilting do not project strongly onto the time tendency of relative vorticity – in other words, they do not contribute to the propagation of vorticity anomalies anywhere in the troposphere – though they may play a role in their growth and maintenance, because of their approximate collocation with vorticity anomalies themselves.

4) ANOMALOUS COLUMN-INTEGRATED INTERNAL ENERGY BUDGET

Models of transient disturbances in the tropics can often be simplified by assuming a first-baroclinic mode vertical structure. This allows one to effectively remove the vertical dimension from the problem and construct a two-dimensional shallow-water-like models, where the relevant internal energy and moisture budgets are vertically-integrated (e.g. Neelin and Zeng 2000; Adames and Kim 2015; Adames and Ming 2018a). To understand the importance of terms in these anomalous budgets, particularly in the context of these simplified models, we therefore look at their vertically-integrated forms.

Following Neelin (2007), the terms in the anomalous vertically-integrated internal energy equation are written below:

$$C_p \frac{\partial \{T\}'}{\partial t} = -C_p \{\mathbf{v} \cdot \nabla T\}' - \left\{ \omega \frac{\partial s}{\partial p} \right\}' + P' + F' + H'. \quad (12)$$

Here C_p is the specific heat of dry air at constant pressure; T is the temperature; $s = C_p T + gz$ is the dry static energy; P is heating due to condensation of water vapor associated with precipitation; F is the net column radiation; and H is the sensible heat flux. The curly braces signify mass-weighted integration over the full column of the quantity inside:

$$\{(\cdot)\} = \frac{1}{g} \int_0^{P_s} (\cdot) dp. \quad (13)$$

Ultimately, the net radiation and sensible heat terms in the anomalous budget make negligible contributions to the total; therefore we plot the anomalous terms of the following approximate form of the budget:

$$C_p \frac{\partial \{T\}'}{\partial t} \approx -C_p \{\mathbf{v} \cdot \nabla T\}' - \left\{ \omega \frac{\partial s}{\partial p} \right\}' + P', \quad (14)$$

which is exactly the same as Equation 12 with the exception of our ignoring of F' and H' . For a comprehensive plot of the spatial pattern of the all of the anomalous terms in the column-integrated internal energy budget, including F' and H' , as well as a Reynolds decomposition of the horizontal advection term, performed following the methodology used for the EPV budget, see Figure S5 in the supplement.

The terms in the approximate budget are plotted in Figure 11 along with contours indicating the values of anomalous vertically-integrated internal energy, $C_p \{T\}'$. In Figure 11(a) we can see a negative anomaly in internal energy at the storm center, flanked by an anomalous negative internal energy tendency to the west and an anomalous positive internal energy tendency to the east; this dipole pattern in the tendency is consistent with the westward propagation of the negative internal energy anomaly at the storm center. The term on the right hand side of the budget that projects most strongly onto the time tendency is the sum of the vertical advection of dry static energy and the column-integrated latent heating associated with precipitation [Figure 11(c)]; overall this has a projection value of 2.65 on the tendency over the domain plotted. Horizontal advection of internal energy serves to damp this propagation tendency [Figure 11(b)]. Total horizontal advection has a projection value of -1.45 . Of this damping influence, horizontal advection of low-frequency internal energy by the high-frequency meridional wind, $-\frac{C_p}{a} \left\{ v' \frac{\partial \bar{T}}{\partial \phi} \right\}'$, and horizontal advection of the high-frequency internal energy by the low-frequency zonal wind, $-\frac{C_p}{a \cos \phi} \left\{ \bar{u} \frac{\partial T'}{\partial \lambda} \right\}'$, contribute -1.18 and -0.16 to the projection, respectively, indicating that the horizontal advection term is primarily due to the high-frequency meridional wind acting on the low-frequency merid-

ional temperature gradient. The low-frequency meridional temperature gradient is positive due to the imposed the land-ocean contrast in heat capacity, though slightly weaker than the observed meridional temperature gradient as shown in Figure 1(e) and (f). The small contribution of the $-\frac{C_p}{a \cos \phi} \left\{ \bar{u} \frac{\partial T'}{\partial \lambda} \right\}'$ term is due to the westward advection of the upper-level temperature anomalies by the low-frequency upper-level easterly winds, e.g. in Figure 5. The residual term is plotted in Figure 11(d); there it is clear that while the anomalies do project negatively onto the spatial pattern of the time tendency, the magnitudes of the anomalies are small. A full tabulation of the projections of each term in the decomposed internal energy budget (Equation 12) can be found in Figure 12.

5) ANOMALOUS COLUMN-INTEGRATED MOISTURE BUDGET

Following Adames and Ming (2018a), the anomalous column-integrated moisture budget can be written as:

$$L_v \frac{\partial \{q_v\}'}{\partial t} = -L_v \{\mathbf{v} \cdot \nabla q_v\}' - L_v \left\{ \omega \frac{\partial q_v}{\partial p} \right\}' - P' + E'. \quad (15)$$

Here q_v represents the specific humidity and P' and E' represent the precipitation and evaporation, respectively, each implicitly scaled by L_v , the latent heat of vaporization, to have units of W m^{-2} to be consistent with the convention used in the internal energy equation. The theory of Adames and Ming (2018b) assumes that of the terms in the anomalous budget in Equation 15, only the horizontal advection of the low-frequency moisture by the high-frequency meridional winds, vertical advection of moisture, and precipitation anomalies are important. It is worth verifying whether this is true in our simulation.

Again, we compute the terms in the anomalous budget following the methods described in Section 2c.3. The results are shown in Figure 13. The time tendency anomaly pattern, panel (a), depicts an east-west-oriented dipole pattern, consistent with the westward propagation of the storms.

647 The two largest terms on the right hand side of the budget are the vertical advection and precip-
648 itation terms, each with maximum magnitudes on the order of 60 W m^{-2} , with vertical advection
649 being a net source (positive) and precipitation being a net sink (negative). Since they largely offset
650 each other, as in Adames and Ming (2018a), we combine these into one term and refer to it as the
651 “column moisture process.” This aggregate term projects strongly onto the time tendency, with
652 a projection value of 0.63 over the region plotted, though perhaps has a slightly northwestward
653 orientation compared with the more westward orientation of the tendency itself. Horizontal ad-
654 vection plays a secondary role, and acts to turn the dipole orientation more toward the west (with
655 a projection value of 0.42). In terms of propagation, the anomalous latent heat fluxes, panel (d),
656 play a minor damping role, with a projection of -0.05 ; however they are positive, and collocated
657 mainly with the column moisture anomalies associated with the disturbance, and therefore more
658 directly contribute to their in-place amplification. In the projection sense, these results are largely
659 consistent with the results of Adames and Ming (2018a) in AM4; there the column moisture pro-
660 cess term was dominant, with a minor positive contribution coming from horizontal advection,
661 and a minor negative contribution coming from evaporation.

662 Similar to what we did with the EPV and internal energy budgets, we can decompose the
663 horizontal advection term into components due to the product of the low-frequency winds and
664 high-frequency moisture gradients, products of the high-frequency winds and the low-frequency
665 moisture gradients, and products of the high-frequency factors. This allows us to continue to de-
666 termine the feasibility of using a linear model of the column-integrated moisture equation. The
667 spatial patterns of the terms associated with this decomposition compared with all other terms in
668 the anomalous column-integrated moisture budget can be found in Figure S6 of the supplement.
669 For brevity, we show only the projection of each of these terms on the anomalous time tendency
670 of column-integrated moisture in the main body of this manuscript in Figure 14. Here we find

that the primary reason for the positive contribution of the horizontal advection of moisture to the westward-propagation tendency is the component due to advection of high-frequency moisture anomalies by the low-frequency zonal winds, with a projection of 0.12. This is followed in projection magnitude by the term related to the product of the high-frequency meridional wind and high-frequency meridional moisture gradient, 0.08, and the term related to the product of the low-frequency meridional wind and high-frequency meridional moisture gradient, 0.07. We will note, however, that the term nonlinear in high-frequency factors is not statistically significant over a large portion of the South Asian monsoon region [see Figure S6(h)]. The term related to the product of the high-frequency meridional winds and low-frequency meridional moisture gradient has a strong spatial pattern relative to the other terms that comprise the total horizontal advection term [Figure S6(g)]; however, it does not project strongly onto the anomalous tendency of column-integrated moisture. This could be due to the fact that the MLPSs in our idealized simulation are centered at roughly the local maximum in the JJAS-mean column-integrated moisture field, which coupled with a cyclonic circulation leads to the quadrupole pattern seen in Figure S6(g).

4. Discussion

We can assess the potential applicability of various theories that have appeared in the literature for the growth and propagation of MLPSs to the storms we find in our idealized simulation by interpreting the structure and budget analysis results presented above.

a. Advection by the mean upper-level easterly winds, modified by beta drift

We will begin by discussing one possible propagation mechanism, inspired by the notion that MLPSs could be analogous to tropical depressions (Boos et al. 2015; Cohen and Boos 2016). Here it is suggested that MLPSs could be steered westward by the low-frequency upper-level

693 easterly winds, and have an additional northward component to their motion through beta drift,
694 i.e. nonlinear advection of high-frequency vorticity anomalies by the high-frequency winds. Boos
695 et al. (2015) base this hypothesis off of a composite analysis of South Asian monsoon depressions
696 using tracks and positions from their own archive (Hurley and Boos 2015) and meteorological
697 variables derived from the ERA-Interim reanalysis (Dee et al. 2011). In looking at the structure of
698 EPV anomalies of the storms in our simulation in the pressure versus latitude plane, with the JJAS
699 mean zonal winds overlaid, Figure 5(b), we find that steering by the mean upper-level easterly
700 winds is likely not the case for our disturbances. The majority of the positive EPV anomalies are
701 located in a region of mean westerly winds, and where there is overlap with easterly winds, the
702 winds are too weak to propagate the vortex westward at 6 m s^{-1} .

703 Beta drift is indicative of nonlinear advection of high-frequency EPV anomalies by high-
704 frequency wind anomalies and is normally responsible for a meridional component to a storm's
705 path (Tamarin and Kaspi 2016), which indeed is significant for depressions seen in observations
706 (Hurley and Boos 2015; Boos et al. 2015). In our analysis, we find that there is little nonlinear
707 contribution to the anomalous horizontal advection term in the EPV budget, which suggests that
708 beta drift is not playing a major role in the propagation of the storms either. This is consistent with
709 the fact that the storms have little northward component to their propagation, which is a deviation
710 from those in observations (Hurley and Boos 2015).

711 *b. Baroclinic instability*

712 Some of the original theories for MLPSs were based in the idea that they emerged out of dry or
713 moist baroclinic instability in the presence of an easterly vertical wind shear (Mishra and Salvekar
714 1980; Mak 1983; Moorthi and Arakawa 1985). Indeed this type of mechanism has been used to
715 explain MLPS-like disturbances that occurred in a similarly idealized model roughly twenty years

ago (Xie and Saiki 1999). Cohen and Boos (2016) argue, however, based on the structure of EPV anomalies in the pressure-longitude plane, that moist baroclinic instability is not occurring in observed MLPs. They argue that in order for counterpropagating Rossby waves to interact constructively, their phasing must be such that EPV anomalies tilt against the mean vertical wind shear with height; in other words, EPV anomalies would need to tilt to the east in the South Asian monsoon region with its characteristic mean easterly shear. This is not the case in observations (Cohen and Boos 2016), and it is also not the case for the storms in our idealized simulation.

We find that the structure of EPV and temperature anomalies shown in Figure 5(a) resembles that found for monsoon depressions in Cohen and Boos (2016). There, for storms mid-way through their lifetimes, Cohen and Boos (2016) found that monsoon depressions can be characterized by a positive column of EPV, tilting slightly westward with height, i.e. with the shear, and a warm-over-cold temperature anomaly structure. They contrast these anomaly patterns with those seen for extratropical diabatic Rossby waves, noting in particular that diabatic Rossby waves have EPV anomalies that tilt against the mean wind shear, a necessary condition for growth out of baroclinic instability. Because the positive EPV anomalies in our idealized simulation tilt with the shear¹², we take Figure 5(a) as tentative evidence that baroclinic instability is not playing a role in the life cycle of the low pressure systems simulated in our idealized model.

¹²Above the 200 hPa pressure level there is a strong positive EPV anomaly slightly the east of the mid-tropospheric EPV anomalies, and a negative anomaly directly above the mid-troposphere maximum. It is possible one could interpret these as evidence of tilting against the shear, and thus possible baroclinic instability, following the argument of Cohen and Boos (2016). However, while not apparent in Figure 5(a) due to the statistical significance mask, these anomalies appear rather to be components of a more organized packet of gravity waves propagating upward from the monsoon low pressure system, similar to those seen emanating from convectively coupled waves near the equator (Kiladis et al. 2009), and not part of the low pressure system itself.

733 *c. Moisture vortex instability*

734 Despite baroclinic instability appearing not to play a role in the disturbances in our idealized
735 simulation, their propagation, at least at upper levels, appears to be consistent with the propa-
736 gation mechanism of Rossby waves, i.e. advection of the planetary vorticity by the anomalous
737 meridional winds. At lower levels, analysis of the anomalous EPV budget suggests that prop-
738 agation is driven by latent heat release due to the condensation of water vapor associated with
739 precipitation. Superficially, this is consistent with another recently proposed theory for MLPSs,
740 called “moisture vortex instability” (MVI) (Adames and Ming 2018a).

741 Moisture vortex instability theory is based on using vertically-truncated versions of the momen-
742 tum, thermodynamic, and moisture equations; in this context “vertically-truncated” means that
743 the horizontal winds, temperature, geopotential, and specific humidity are projected onto basis
744 functions consistent with a first-baroclinic mode vertical structure for the vertical velocity. This
745 reduces the equations to a shallow water-like system, which is more amenable to analysis [e.g. as
746 in Neelin and Zeng (2000), Haertel et al. (2008), or Adames and Kim (2015)]. In Adames and
747 Ming (2018b), the truncated equations are linearized about a South Asian monsoon-season-like
748 basic state, and through analysis of a dispersion relation, are shown to support unstable modes.
749 The instability is associated with a partially in-phase relationship between precipitation anomalies
750 (corresponding with upward vertical motion and convergence of low-level horizontal winds) and
751 cyclonic (i.e. positive) vorticity anomalies. The precipitation anomalies, through their associa-
752 tion with low-level convergence, result in a growing tendency for the vorticity anomalies through
753 vortex-stretching (Adames and Ming 2018b). Propagation of the wave in their framework is pri-
754 marily due to vortex stretching from moist convection in regions of isentropic ascent and horizontal
755 moisture advection.

756 In terms of the primitive equations, moisture vortex instability theory depends on the advection
 757 of planetary vorticity, vortex stretching, meridional and vertical advection of the mean internal
 758 energy and moisture by the anomalous winds and latent heating due to precipitation. It assumes
 759 no influence by the mean state winds or terms nonlinear in anomalies (Adames and Ming 2018a).
 760 Our analysis of the anomalous vorticity, column-integrated internal energy, and column-integrated
 761 moisture budgets can shed light on whether the assumptions made in constructing the theory hold
 762 in our idealized simulation.

763 In the case of the vorticity budget, we find that at both 850 hPa and 400 hPa that propagation is
 764 dominated by the terms associated with advection of planetary vorticity and vortex stretching, i.e.
 765 the term $-\nabla \cdot (f\mathbf{u})'$, which is consistent with MVI theory. The term related to the influence of the
 766 collective influence of vertical vorticity advection and vortex tilting, i.e. $-\nabla \cdot \left(\omega \hat{\mathbf{k}} \times \frac{\partial \mathbf{u}}{\partial p} \right)'$ does
 767 project strongly onto relative vorticity anomalies at 400 hPa, however, which suggests that it plays
 768 a role in the growth and maintenance of vorticity of the MLPSs in our simulation, which is not
 769 consistent with MVI. This term happens to be dominated by the contribution of the divergence of
 770 the product between the anomalous vertical velocity and the low-frequency vertical shear of the
 771 winds (not shown), indicating the potential importance of including the low-frequency winds in a
 772 theory for MLPSs.

773 In the case of the column-integrated internal energy budget, the picture of the storms in our sim-
 774 ulation is largely consistent with the assumptions of MVI theory. There the anomalous radiative
 775 and sensible heating parts of the thermodynamic equation were neglected, and they are indeed
 776 found to be negligible in our simulation. The terms retained in the anomalous thermodynamic
 777 budget in Adames and Ming (2018a) were the vertical advection of mean dry static energy by the
 778 anomalous pressure velocity, the column latent heating due to precipitation, and meridional ad-

vection of mean internal energy by the anomalous meridional wind. These are indeed the leading order terms in the anomalous thermodynamic budget in our simulation (see Figure 12).

Finally, as assumed in Adames and Ming (2018b), the vertical advection of moisture and the loss of column moisture through precipitation play an important role in the moisture budget. That said, assumptions made regarding the horizontal advection of moisture in Adames and Ming (2018b) do not necessarily hold in our simulation. Adames and Ming (2018b) assume that advection of mean moisture by the anomalous meridional wind plays a leading-order role in the budget. We find that this does not quite hold in our simulation. While the advection of low-frequency moisture by the high-frequency meridional wind anomalies has a relatively strong magnitude relative to other terms related to horizontal advection, it does not project strongly onto the anomalous time tendency of column-integrated moisture. This is likely in part because the storms we analyze in this simulation are centered roughly at the latitude of maximum JJAS-mean column-integrated moisture, in contrast to the assumption made by Adames and Ming (2018a), that the storms form in an area of uniformly increasing column moisture in the meridional direction. A term that projects more strongly is the advection of the high-frequency moisture anomalies by the low-frequency zonal wind, which was not considered in Adames and Ming (2018a).

5. Conclusion

In this study we have completed a systematic analysis of low pressure systems in the South Asian monsoon region in a heavily-idealized moist GCM, notably without impacts of clouds or topography. The low pressure systems found in our simulation share a number of characteristics with South Asian monsoon low pressure systems observed in reality and those simulated in GFDL's AM4. For example precipitation anomalies in the South Asian monsoon region in our simulation have a typical zonal scale of around zonal wavenumber 10, consistent with the scale

802 seen in TRMM observations and AM4; the typical frequency of around 0.2 d^{-1} is consistent with
803 that found in those datasets as well (Adames and Ming 2018a). In addition, we find that the verti-
804 cal structure of potential vorticity anomalies associated with the low pressure systems simulated in
805 our model shares an important qualitative feature with that found in reanalysis: the EPV anoma-
806 lies in the troposphere tilt slightly with the JJAS mean easterly zonal wind shear (Cohen and Boos
807 2016).

808 Aspects of the low pressure systems that differ slightly from those seen in reality are their prop-
809 agation speed and direction. In our simulation, the storms propagate predominantly westward at
810 speeds of over 6 ms^{-1} ; this is faster than storms seen in GFDL's AM4 and reanalysis. There are
811 several possible explanations for this difference. Two of these arise from Rossby wave theory.
812 From inspection of Figure 4 it is possible that these waves are of slightly larger scale, resulting in
813 a smaller horizontal wavenumber, than the low pressure systems simulated in AM4 and observed
814 in reanalysis. In addition, because these systems occur at a lower latitude than in the aforemen-
815 tioned datasets, the Rossby radius of deformation is larger, which would also cause these systems
816 to exhibit faster westward propagation [see Equation 22a in Adames and Ming (2018b)]. We find
817 little meridional component to the propagation direction, which is different than at least reanalysis
818 Boos et al. (2015); in GFDL's AM4 model, storms propagated predominantly westward as well. It
819 is possible that the northward component of propagation is largely a result of nonlinear beta drift,
820 which is characteristic of the stronger storms that were analyzed by Boos et al. (2015). Per the
821 role of the meridional temperature gradient in the propagation of column-integrated moist enthalpy
822 anomalies associated with MLPSs suggested by moisture vortex instability theory (Adames and
823 Ming 2018a), the more equatorward location of the systems when compared with observations
824 may be in part due to the more equatorward meridional temperature gradient in our simulation.

825 The movement of the weak disturbances in our simulation can largely be explained through
826 linearized versions of the primitive equations, rather than beta drift, as was the case for monsoon
827 depressions analyzed in reanalysis in Boos et al. (2015). A slight deviation from moisture vortex
828 instability theory is the potential contribution of the vertical shear of the low-frequency winds to
829 the growth and maintenance of the disturbances. Despite not being classic examples of baroclinic
830 instability, these disturbances might still be classified as a certain form of diabatically-influenced
831 Rossby wave, like that described in Adames and Ming (2018b). A major distinguishing factor
832 between extratropical diabatic Rossby waves and Rossby-like waves deriving from moisture vortex
833 instability is that, while influenced by moist processes, extratropical Rossby waves can grow in
834 the absence of precipitation (Vallis 2006); for moisture vortex instability, precipitation is necessary
835 for growth (Adames and Ming 2018b). The possibility of an approximate explanation via a linear
836 model could motivate further sensitivity studies in a modeling framework similar to the one used
837 here, to test whether properties of the mean state, like the meridional temperature or moisture
838 gradient, could influence properties of the low pressure systems, like the phase speed.

839 While the work we have done here demonstrates that somewhat realistic MLPS-like disturbances
840 can be simulated with simplified model physics and boundary conditions, it does not rule out that
841 even further idealizations could be made. We intentionally used realistic continental geometry,
842 reduced heat capacity, and a bucket hydrology model to limit evaporation over land, as to remove
843 those as possible reasons for too unrealistic a mean climate to support MLPSs. In future work it
844 could be useful to run simulations with realistic continental geometry, with only reduced heat ca-
845 pacity over land, or only bucket hydrology over land to see which is most important in generating
846 a mean state climate suitable for South Asian MLPSs. For example, this could inform us whether
847 it might be valuable to use a simpler land setup such as the “moist land” simulations with a rectan-
848 gular continent in Zhou and Xie (2018) when attempting to systematically modify the mean state

as suggested at the end of the previous paragraph. In those simulations in Zhou and Xie (2018), the continental geometry is significantly simplified, and the land surface is assumed to always be saturated; that is, the only thing distinguishing land from ocean is its reduced heat capacity.

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APPENDIX

Method for computing the internal energy budget

To compute the terms in Equation 12, we start from the dry static energy budget, which has a similar form:

$$\frac{\partial \{s\}}{\partial t} = -\{\mathbf{v} \cdot \nabla s\} - \left\{ \omega \frac{\partial s}{\partial p} \right\} + P + F + H. \quad (\text{A1})$$

Unlike Equation 12, however, the advection terms can be placed in flux form:

$$\frac{\partial \{s\}}{\partial t} = -\nabla \cdot \{\mathbf{v}s\} + P + F + H. \quad (\text{A2})$$

With the budget in this form, we can apply the procedure outlined in Appendix A of Hill et al. (2017) to compute a barotropic adjustment, $\mathbf{v}_{\text{adj}}$, to the horizontal winds such that the budget in Equation A2 is closed. The adjusted horizontal winds are thus:

$$\mathbf{v} = \mathbf{v}_{\text{raw}} - \mathbf{v}_{\text{adj}}, \quad (\text{A3})$$

867 where $\mathbf{v}_{\text{raw}}$ are the raw horizontal winds output on model-native vertical levels. Using the adjusted
 868 winds to compute the horizontal advection term in Equation A1 explicitly, we can then compute
 869 the vertical advection term as a residual:

$$-\left\{\omega \frac{\partial s}{\partial p}\right\} = \frac{\partial \{s\}}{\partial t} + \{\mathbf{v} \cdot \nabla s\} - P - F - H, \quad (\text{A4})$$

870 where the terms on the right hand side are all computed explicitly. Note that this is the same
 871 vertical advection term as in internal energy budget, Equation 12. Accordingly, we can use the
 872 adjusted winds derived here to compute the horizontal temperature advection terms in Equation 12:

$$C_p \frac{\partial \{T\}'}{\partial t} = -C_p \{\mathbf{v} \cdot \nabla T\}' - \left\{\omega \frac{\partial s}{\partial p}\right\}' + P' + F' + H'. \quad (\text{A5})$$

873 Because we compute the vertical advection term as a residual from the dry static energy budget,
 874 when we plug it back in to the internal energy equation, Equation A5, we have a small residual,
 875 shown in Figure 11(d).

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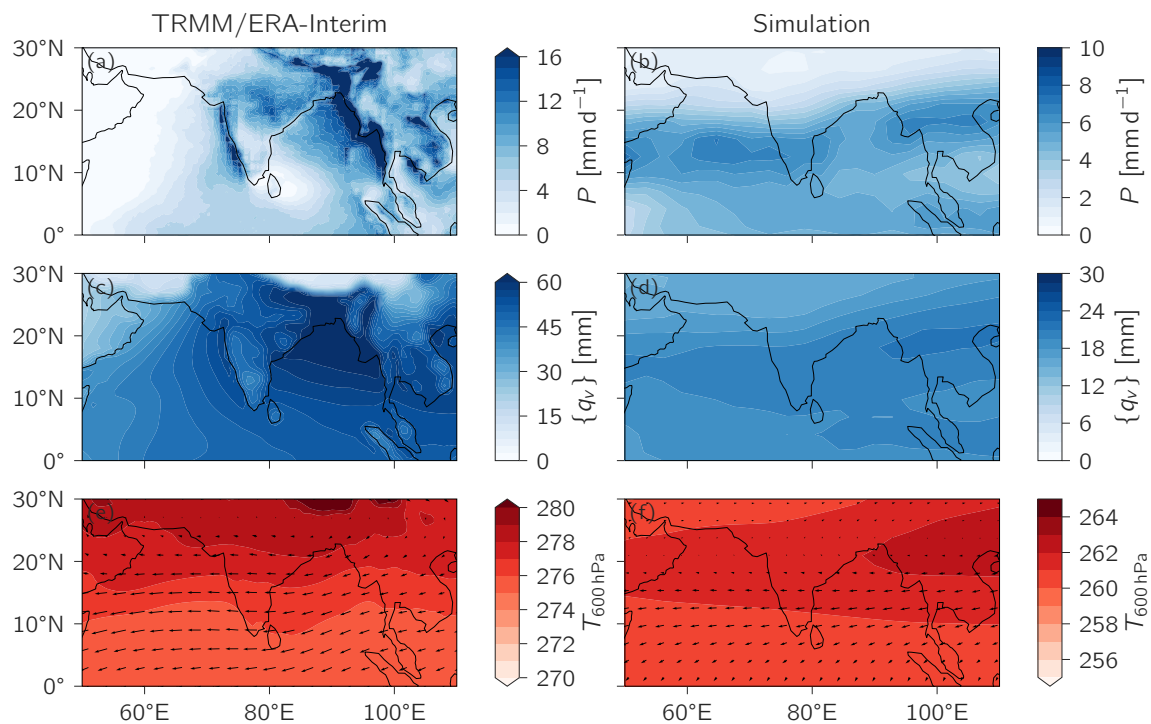


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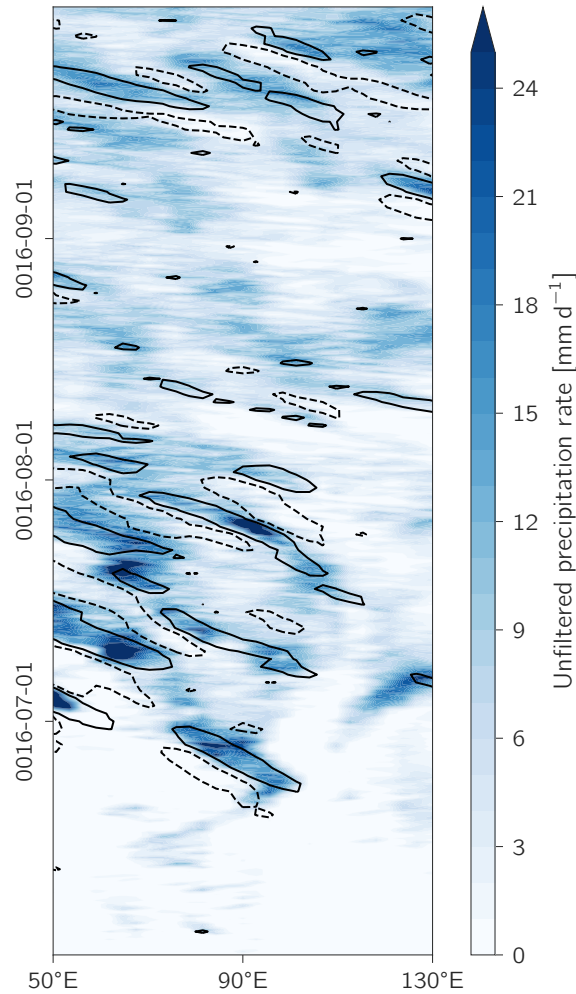


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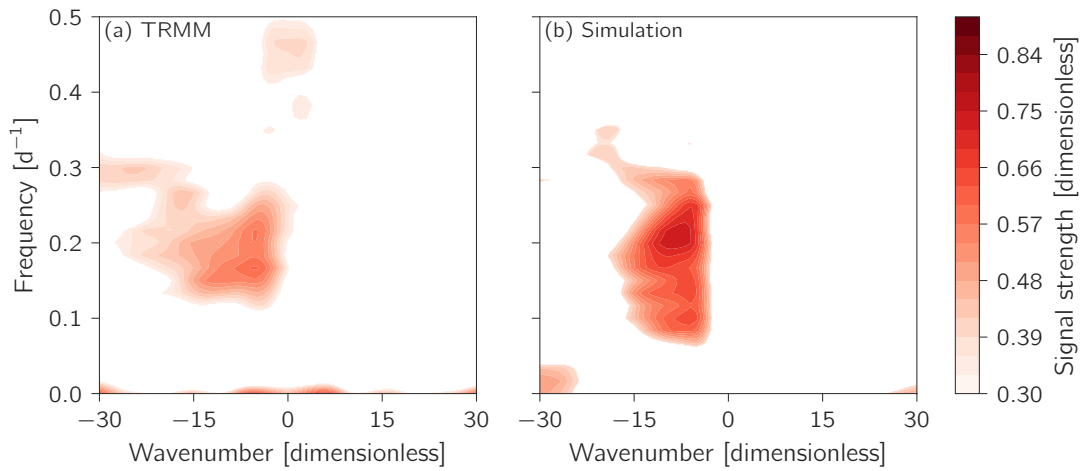


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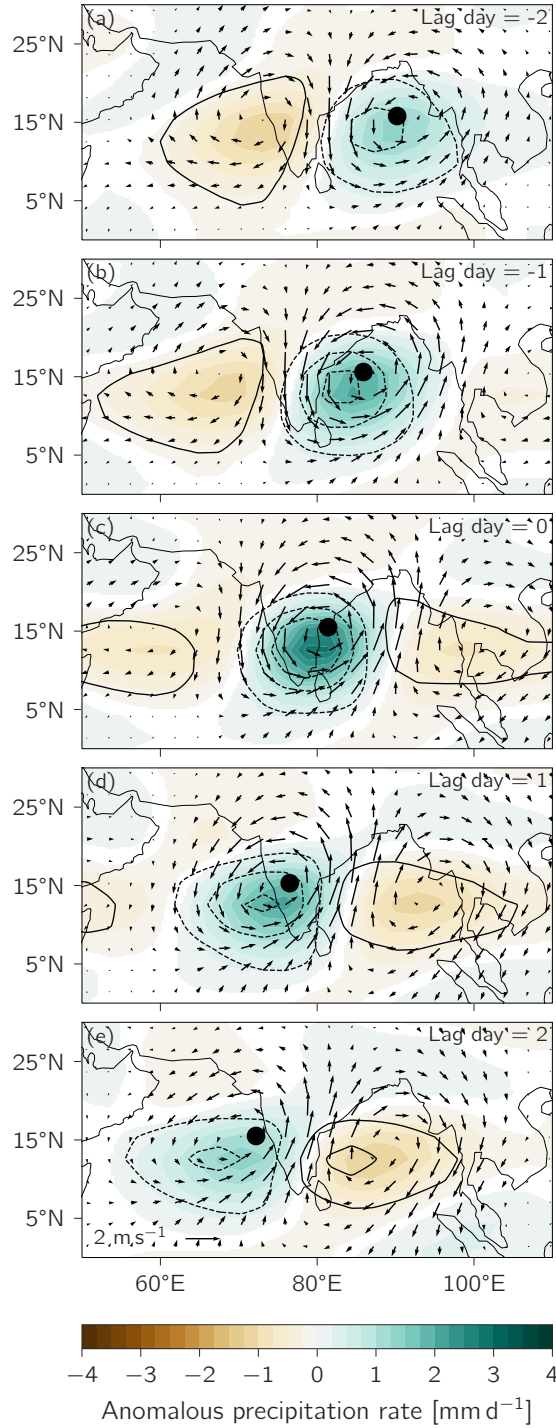


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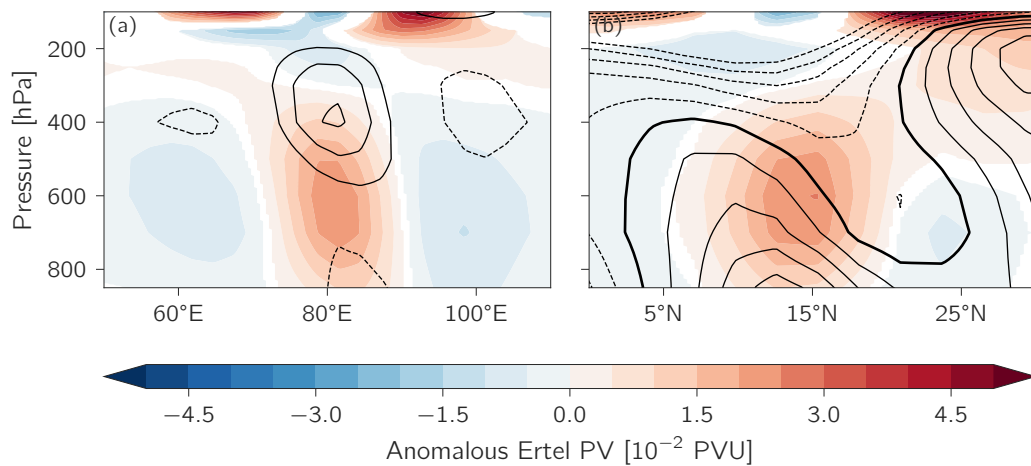


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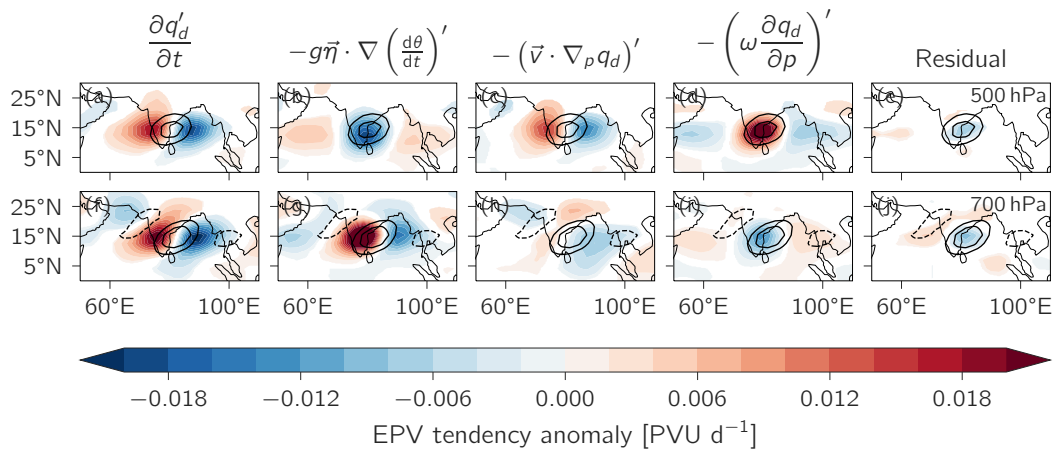


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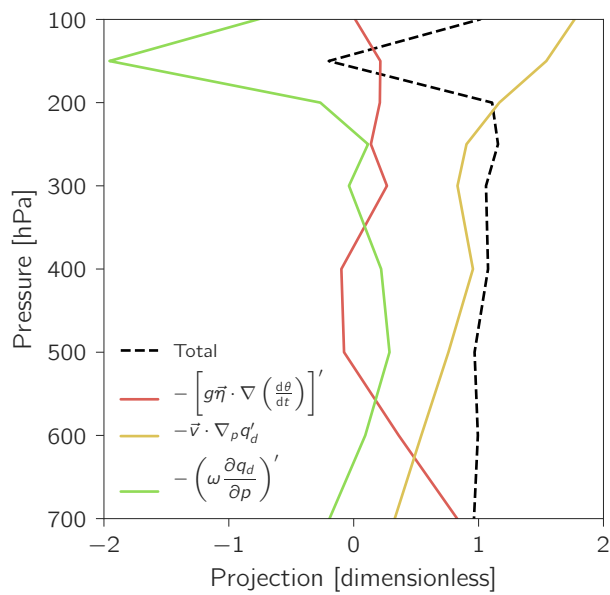


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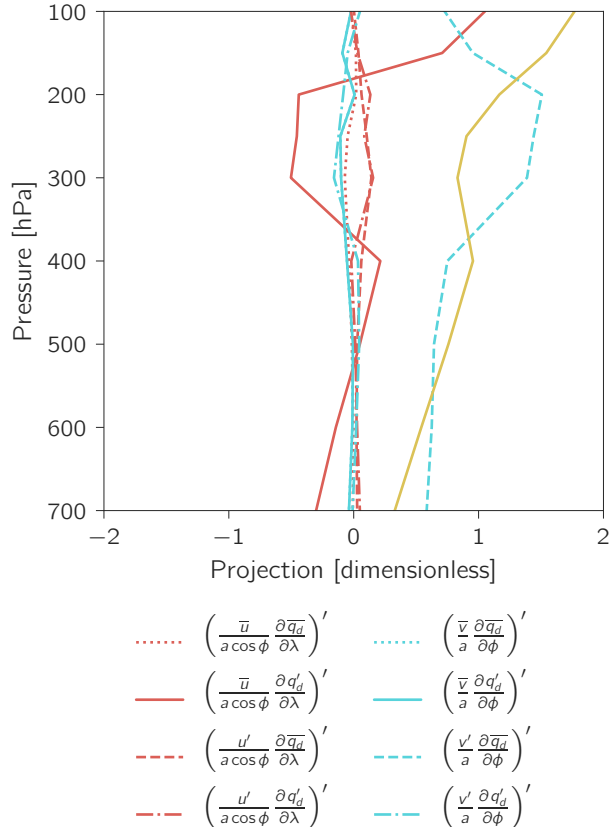


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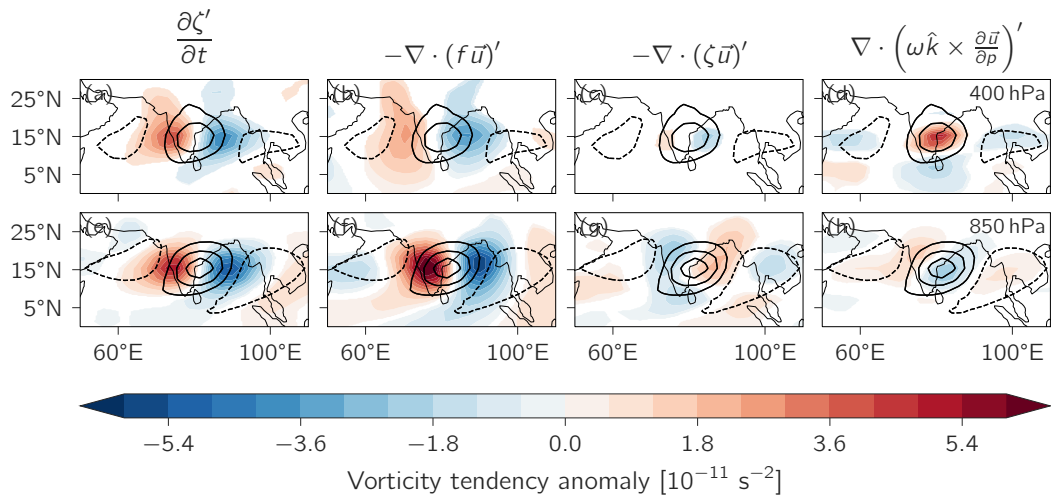


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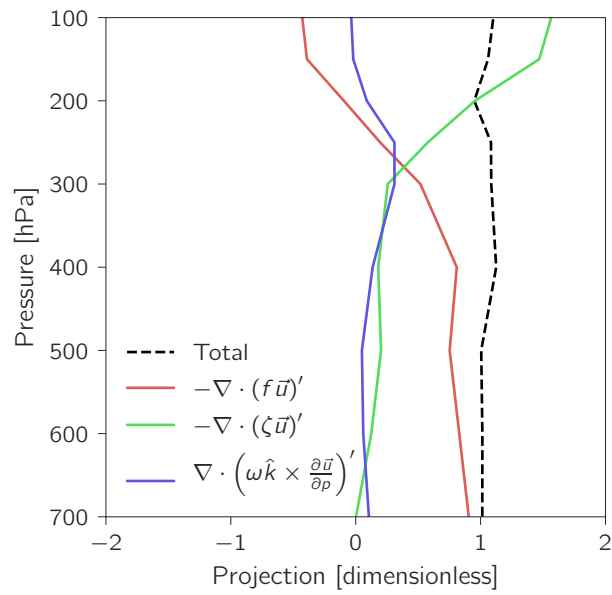


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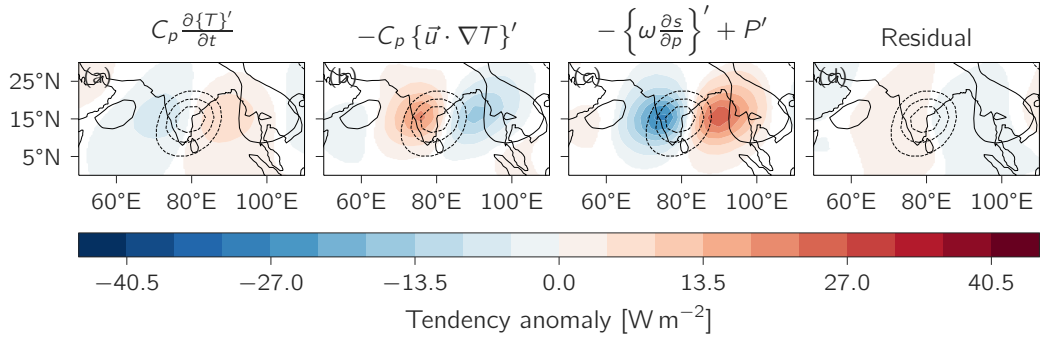


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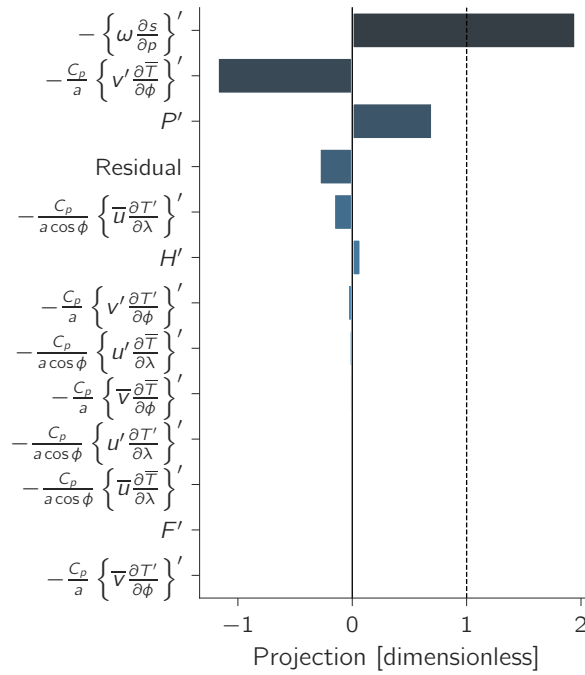


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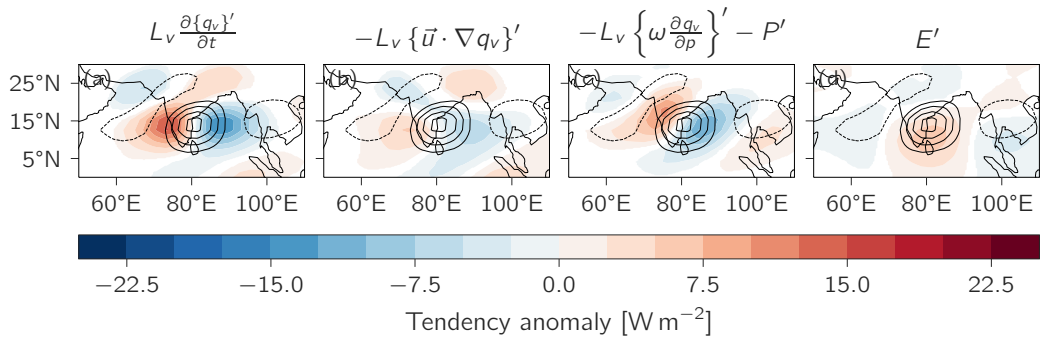


FIG. 13. Terms in the anomalous moisture budget (colors). Only tendency anomalies statistically significant at the 99% level are shown. Contours represent column-integrated moisture anomalies scaled by the latent heat of vaporization, L_v . Negative contours are dashed; positive contours are solid. With the exception of the omission of the zero contour, contours are separated by intervals of $4.0 \times 10^5 \text{ J m}^{-2}$.

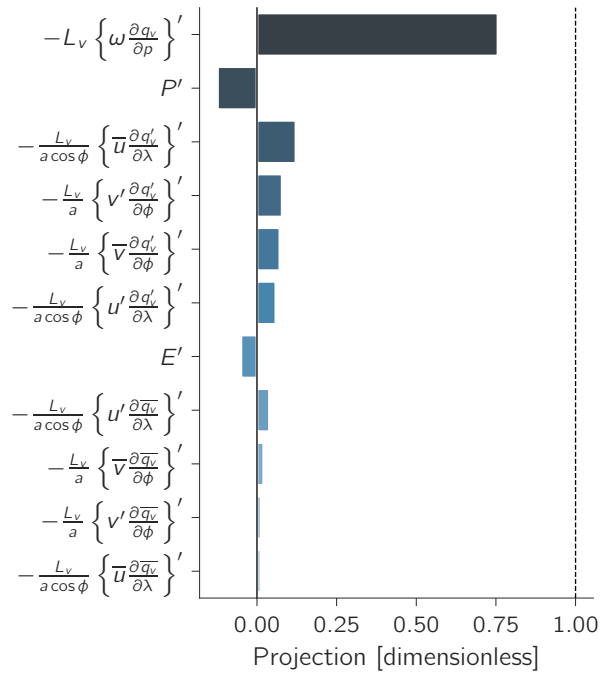


FIG. 14. Projection of the terms on the right hand side of the column-integrated moisture budget sorted in descending order by absolute value. The horizontal advection term is comprehensively decomposed into linear and nonlinear components. The sum of the components adds exactly to 1.0.