	Status - INDUSTRY - COMMENT
1	A Case Study of the Physical Processes Associated with the Atmospheric River Initial Condition
2	Sensitivity from an Adjoint Model
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Abstract

25	Analysis of a strong landfalling atmospheric river is presented that compares the
26	evolution of a control simulation with that of an adjoint-derived perturbed simulation using the
27	Coupled Ocean/Atmosphere Mesoscale Prediction System. The initial condition sensitivities are
28	optimized for all state variables to maximize the accumulated precipitation within the majority of
29	California. The water vapor transport is found to be substantially enhanced at the California
30	coast in the perturbed simulation during the time of peak precipitation, demonstrating a
31	strengthened role of the orographic precipitation forcing. Similarly, moisture convergence and
32	vertical velocities derived from the transverse circulation are found to be substantially enhanced
33	during the time of peak precipitation, also demonstrating a strengthened role of the dynamic
34	component of the precipitation.
35	Importantly, both components of precipitation are associated with enhanced latent
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46 1. Introduction

47 The predictability of atmospheric rivers (ARs) is vitally important for the management of water resources in California (CA) since it is a densely populated region and highly dependent 48 49 on the arrival, or lack thereof, of just a handful of precipitation events per year (Dettinger 2013; Ralph et al. 2013; Dettinger and Cayan 2014; Ralph 2017; Lamjiri et al. 2017, 2018). DeFlorio et 50 51 al. (2018) performed an assessment of global ARs and found that at a 2-day lead time, fewer than 52 75% of the European Centre for Medium-Range Weather Forecasts (ECMWF) ensemble 53 members predicted ARs within 250 km of the actual landfall location, thus motivating the need 54 for forecast improvements in not only the magnitude of the water vapor content in the AR, but also its position. The demand from California water managers for more accurate predictions of 55 56 both AR strength and landfall position has focused attention on improving numerical prediction systems, as well as aspects of the initial conditions that impact AR characteristics. A version of 57 the Weather Research and Forecasting (WRF) model tailored for the needs of the U.S. West 58 59 Coast and employed for real time weather forecasts by Martin et al. (2019) demonstrated improved forecast skill over the Global Forecast System (GFS) for 15 landfalling events in CA. 60 Improvements in the initial conditions are being made in part through the AR Reconnaissance 61 62 (AR-Recon) program which targets ARs using airborne instruments, having an emphasis on the in situ dropsondes (Ralph 2018). The optimal aircraft tracks and dropsondes locations, however, 63 64 are typically a matter for debate because it is not always clear where additional observations are 65 most likely to improve the analyses and subsequent forecasts. This study focuses on improving the dynamical understanding of how perturbations identified using an adjoint technique grow 66 rapidly and impact AR forecasts. 67

It has long been understood that small errors in the initial conditions can grow rapidly 68 due to the chaotic nature of the atmosphere (Lorenz 1969). However, an initial condition error of 69 a state variable can have a range of downstream impacts depending on how it projects onto fast 70 growing perturbations (e.g., Reynolds et al. 2001, 2019; Doyle et al. 2014). Atmospheric 71 72 instabilities are the vehicles through which small errors in the initial state may grow rapidly, 73 thereby contributing towards forecast errors. Aircraft reconnaissance is often utilized to reduce the initial condition errors during impactful weather phenomena by collecting in-situ 74 observations of critical regions where there may otherwise be gaps in the observing systems 75 76 (Shoemaker et al. 1990; Gray et al. 1991; Martin and Gray 1993; Majumdar 2016). However, the optimal sampling locations providing the greatest improvement to the forecast are often not 77 obvious. 78

The Naval Research Laboratory (NRL) Coupled Ocean-Atmosphere Mesoscale 79 Prediction system (COAMPS[®]) including its moist adjoint model (Amerault et al. 2008; Doyle et 80 al. 2012, 2014) was run in real time during AR-Recon in 2018 and 2019 to provide an objective 81 82 method for determining where short-term forecasts of precipitation over the west coast of North 83 America are most sensitive to changes in the initial state. Although not always the case, the 84 regions highlighted by the moist adjoint tool are often broadly consistent with the forecaster's dynamical understanding of phenomena that would impact the forecast evolution. By comparing 85 dropsondes with short-range forecasts, Lavers et al. (2018) found relative errors of about 20% in 86 87 the water vapor transport within ARs present in the ECMWF forecast prediction system. Since 88 water vapor transport is highly correlated with orographic precipitation (Alpert 1986; Neiman et 89 al. 2002, 2009a; Lu et al. 2010; Ralph et al. 2013), this suggests that aircraft reconnaissance sampling should be focused in regions of strongest water vapor transport. However, there is 90

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91 evidence that multiple dynamical features on a variety scales may have comparable impact on precipitation forecasts. For example, Zhang et al. (2019) found that ARs are linked to a parent 92 extratropical cyclone (EC) 80% of the time, suggesting that increased observations, and more 93 accurate analyses, of the dynamically-active regions associated with the development of the EC 94 95 are also important (Zheng et al. 2013). A dynamical investigation of the sensitivity results and 96 subsequent perturbation growth will both increase understanding of the physical mechanisms that modify AR evolution, and provide context for the use of sensitivity tools for AR-97 Reconnaissance observing strategies. 98

The present study is motivated by, and complimentary to, Reynolds et al. (2019), which 99 100 used the NRL moist adjoint modeling system to evaluate the sensitivity of AR forecasts for January and February 2017. They found that both the accumulated precipitation and the low-101 102 level kinetic energy forecast were most sensitive to errors in and around the AR. Furthermore, 103 they showed that the largest forecast sensitivities are from the AR moisture content, followed by temperature and winds, consistent with the results of Doyle et al. (2014) and Doyle et al. (2019) 104 105 for Atlantic ECs. Finally, they demonstrated that the strength of the initial condition sensitivities correlates well with forecast errors, leading the authors to conclude that the moist adjoint is 106 107 relevant for predictability and targeted observation applications.

108 The goal of the present study is to investigate the physical processes responsible for the 109 rapid growth of perturbations derived from a moist adjoint model for an extreme landfalling AR. 110 Out of many different landfalling AR candidates, the 7–9 January 2017 event is investigated 111 because it occurred during the record-breaking precipitation year for Northern California and 112 contained several interesting dynamical features, including a mesoscale frontal wave (Bjerkness

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113	and Solberg	(1922); Renfrew et al	. 1996, Parker 199	7, Rivals et al.	1997, and Hewson 2009
			/	/	/

114 Neiman et al. 2016; Martin et al. 2019), which is notoriously difficult to accurately forecast.

115

116 **2. Methods**

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a) Control and Adjoint Run Descriptions

The January 7–9, 2017 landfalling AR is simulated using the COAMPS modeling system 118 (Hodur 1997; Doyle et al. 2014). The COAMPS model is a nonlinear, nonhydrostatic, 119 120 compressible, terrain-following numerical model. Our simulation is run on a 221x161 grid with a horizontal grid spacing of 40 km on a Lambert-Conformal grid having 70 vertical levels; output 121 is interpolated onto a pressure grid from 1000 to 200 hPa every 25 hPa and from 200 to 50 hPa 122 every 50 hPa. For a full description of the parameterizations and physics packages, see Doyle et 123 al. (2012; 2019). The relatively coarse resolution of 40 km is chosen because the tangent linear 124 approximation tends to be valid for shorter time periods as resolution becomes finer. Future 125 126 work will consider higher-resolution experiments for shorter forecast intervals.

127 The following paragraphs describing the adjoint model closely follows the summaries given in Reynolds et al. (2016, 2019) and Doyle et al. (2014), and is fully described in Doyle et 128 al. (2012). The COAMPS adjoint and Tangent Linear Models (TLM) are described in detail in 129 Amerault et al. (2008) and Doyle et al. (2012). The same physical parameterizations are used in 130 131 the nonlinear, tangent linear, and adjoint models, including ice-phase microphysics as in (Doyle et al. 2012, 2014, 2019; Reynolds et al. 2019). The longwave and short-wave radiative processes 132 133 are included in the nonlinear forecasts but are ignored in the TLM and adjoint in order to avoid highly nonlinear components of the physics. The COAMPS forecast and adjoint systems have 134

been used extensively in studies ranging from tropical cyclone (TC) predictability studies (Doyle
et al. 2011, 2012; Reynolds et al. 2016), to the 2018 and 2019 AR-Recon field projects (Ralph et
al. 2018; Reynolds et al. 2019).

138 The moist adjoint system represents the sensitivity of a forecast metric J (for model state 139 x_t at time t) to each component of an earlier model state (x_{t0}) as

140
$$J(\mathbf{x}_t) = J[\mathbf{M}_{nl}(\mathbf{x}_{t0})]$$
, (1)

where *J* is the response function and M_{nl} is the nonlinear model (see Errico, 1997 for a description of an adjoint model). The response function selected in this study is the 18-h accumulated precipitation averaged over the majority of California (black box in Fig. 2). The gradient of *J* with respect to the initial model state is given by,

145
$$\frac{\partial J}{\partial x_{t0}} = \boldsymbol{M}_{tlm}^T \frac{\partial J}{\partial x_t} , \qquad (2)$$

146 where M_{tlm} is the tangent linear model of M_{nl} and superscript *T* denotes the transpose. The term 147 $\frac{\partial J}{\partial x_t}$ is computed through differentiation of *J* with respect to the model state at time *t*. A

148 requirement of J is that it must be a differentiable and continuous function.

This study is focused on the comparison of the nonlinear control forecast (CTL) with that of an adjoint-derived optimally perturbed TLM following the methods of Oortwijn and Barkmeijer (1995), Rabier et al. (1996), and Errico and Raeder (1999). These initial-time perturbations based on the adjoint sensitivity are designed to grow rapidly and result in large changes to the forecast response function. The production of these optimal perturbations follows directly from that used in Doyle et al. (2012, 2014). Perturbations to the response function *J* are expressed as

$$J' = \sum_j \frac{\partial J}{\partial x_j} x'_j ,$$

157 where $\frac{\partial J}{\partial x_j}$ is the gradient of the response function with respect to the jth initial state component.

(3)

158 The jth component of the perturbation vector x' is optimal when defined such that

159
$$x'_{j} = \frac{s}{w_{j}} \frac{\partial J}{\partial x_{j}}, \qquad (4)$$

for weights w_i . The method for calculating the weights (w_i) and scaling parameter s is detailed 160 in Doyle et al. (2014). The scaling parameter, s (units of J^{-1}), is selected such that the largest 161 perturbation of the water vapor, potential temperature, or zonal wind speed at the initial time 162 does not exceed 1 g kg⁻¹, 1 K, or 1 m s⁻¹, respectively. These perturbation magnitudes are chosen 163 164 such that they are comparable to, or smaller than, errors assigned to radiosonde and dropsonde observations in the data assimilation system [1 K, 1.8 m s⁻¹ and 10% relative humidity at 925 hPa 165 (~1-1.5 g kg⁻¹)]. The optimal perturbations are calculated for potential temperature, the Exner 166 pressure perturbation, mixing ratio, all wind speed components, and microphysical species. 167

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b) The Sawyer-Eliassen Transverse Circulation Model

The Sawyer-Eliassen framework is a well-known method for calculating the 2D
ageostrophic circulation within a jet-front cross section. Following the methods of (Sawyer 1956;
Eliassen 1962; Keyser and Shapiro 1986; Winters and Martin 2014), the Sawyer-Eliassen
equation is given by,

173
$$\left(-\gamma \frac{\partial \theta}{\partial p}\right) \frac{\partial^2 \Psi}{\partial y^2} + \left(2 \frac{\partial M}{\partial p}\right) \frac{\partial^2 \Psi}{\partial p \partial y} + \left(-\frac{\partial M}{\partial y}\right) \frac{\partial^2 \Psi}{\partial^2 p} = Q_g - \gamma \frac{\partial}{\partial y} \left(\frac{d\theta}{dt}\right),$$
(5)

174 where *y* is in the across-jet direction, *x* is in the along-jet direction,
$$\Psi$$
 is the ageostrophic
175 streamfunction, θ is potential temperature, $M = U_g - fy$ is the absolute momentum, $Q_g =$
176 $2\gamma \left(\frac{\partial U_g}{\partial y}\right) \left(\frac{\partial \theta}{\partial x}\right) + 2\gamma \left(\frac{\partial V_g}{\partial y}\right) \left(\frac{\partial \theta}{\partial y}\right)$ is the quasi-geostrophic forcing term (stretching and shearing
8

177 deformations respectively), U_g is the along-jet geostrophic wind, V_g is the across-jet geostrophic

178 wind,
$$\gamma = \frac{R}{fp_0} \left(\frac{p_0}{p}\right)^{\frac{c_v}{c_p}}$$
, $p_0 = 1000$ hPa, $c_v = 718$ J kg⁻¹ K⁻¹, $c_p = 1004$ J kg⁻¹ K⁻¹, $R = 287$ J kg⁻¹

179 K⁻¹, *f* is the Coriolis parameter, and $\frac{d\theta}{dt}$ is the diabatic heating rate. The coefficients on the LHS 180 of equation (5) represent the static stability, baroclinicity, and inertial stability, respectively. The 181 diabatic heating term is calculated following Emanuel et al. (1987) and Winters and Martin 182 (2014), and is given by,

183
$$\frac{d\theta}{dt} = \omega \left(\frac{\partial\theta}{\partial p} - \frac{\Gamma_m}{\Gamma_d} \frac{\theta}{\theta_e} \frac{\partial\theta_e}{\partial p} \right) \tag{6}$$

where θ_e is the equivalent potential temperature calculated following Bryan (2008), Γ_m is the moist adiabatic lapse rate, and Γ_d is the dry adiabatic lapse rate. Equation (6) was found to be in very good agreement with the direct model output of the latent heating, indicating that the diabatic heating is dominated by latent heating. We use a parametrized form for the diabatic heating because only the instantaneous latent heating rate is available from the model output, which can have some very large heating/cooling gradients and small-scale structure, thereby making it difficult for the numerical routine of equation (5) to converge.

Equation (5) is a second-order elliptic partial differential equation (PDE) which is solved numerically on a grid from 1000 to 100 hPa every 25 hPa in the vertical, and interpolated along a transect of length 2,250 km with a 45-km spacing. It is solved using the method of successive over-relaxation (SOR) in a piecewise manner for the dry (Q_g) and moist $(\frac{d\theta}{dt})$ forcing terms separately. The boundary conditions used are $\Psi = 0$ at all boundaries following Shapiro (1981) and Winters and Martin (2014). All cross sections were manually inspected prior to application

197	of the SOR method to ensure that i) the full jet-front structure is captured, and ii) the boundary
198	conditions were appropriate and did not neglect any important features.

199

c) Air Parcel Trajectories

200 Both forward and backward trajectories are utilized in this study, each calculated from 15-minute output, with 25-hPa vertical spacing, on a latitude-longitude grid interpolated to 201 202 0.375° x 0.275°. The method applied to calculate both the forward and backward trajectories is a three-step process which 1) determines the 3-dimensional wind components of each air parcel, 2) 203 advances the air parcels forward by a timestep, and 3) linearly interpolates to the new locations 204 at which point (1) - (2) are repeated following Wernli and Davies (1997) and Wernli (1997). 205 This method allows for an adaptable and computationally inexpensive method of estimating the 206 parcel trajectories. 207

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209 3. Case Study Background

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a) Synoptic-Scale Overview

The heavy precipitation that fell in California during 7–9 January 2017 was due to a landfalling AR event reaching AR Category 4 (CAT4) out of a scale of 5 (Ralph et al. 2019) having two distinct pulses or waves. At the forecast initialization time (Fig. 1a,f), AR Wave 1 (W1) is located ahead of the synoptic scale cyclone's (145W, 36N) cold front and is thereby already associated with a mature EC having a moderate sea-level pressure (SLP) minimum of about 990 hPa and large integrated vapor transport (IVT) values exceeding 800 kg s⁻¹ m⁻¹. Following the eastward propagation of a shortwave seen in the SLP field, the AR makes landfall

218	in Central California beginning around 12 UTC on 7 January with an IVT plume containing
219	maximum values located offshore upwards of 750 kg s ⁻¹ m ⁻¹ (Fig. 1b,g).
220	A resurgence in the IVT (W2) is apparent in Fig. 1c, and is associated with a developing
221	secondary cyclone (C; Fig. 1h), also termed a mesoscale frontal wave (Bjerkness and Solberg
222	(1922); Renfrew et al. 1996, Parker 1997, Rivals et al. 1997, and Hewson 2009; Neiman et al.
223	2016; Martin et al. 2019). This rapid succession of events is consistent with the idea of AR
224	families (Fish et al. 2019). The mesoscale frontal wave is found to be located not only in the left
225	exit region of the upper level jet (Fig. 1h), which is a region favorable for cyclone development,
226	but also along the cold front where strong latent heating is expected. It is thought that the
227	synergy between the forcings at upper and lower levels may be responsible for the development
228	of the mesoscale frontal wave and its intensification into a secondary cyclone. Twelve hours later
229	(Fig. 1d,i), after substantial development in both the SLP and IVT, W2 makes landfall as an AR
230	CAT4 with maximum IVT values reaching 1000 kg s ⁻¹ m ⁻¹ . By the end of the forecast (Fig. 1e,j),
231	the AR has decayed, and the mesoscale frontal wave has propagated poleward to 130W, 47N
232	towards Vancouver.

The Stage-IV (Lin 2011) gridded observational 48-h accumulated precipitation for this 233 234 case is shown in Fig. 2a and exhibits the greatest impacts in three regions: i) the Northern Sierra Nevada Mountains, with values upwards of 200 mm (~8"), ii) along the Northern California 235 coast, with values also upwards of 200 mm, and iii) the Shasta region, with values slightly under 236 237 200 mm. While the CTL run simulates the distribution and magnitude of the Stage-IV precipitation pattern reasonably well in the Sierra Nevada range (120W, 37N) and central coast 238 there are some clear biases in the Shasta (124W, 42N) and north coastal regions which results in 239 only a modestly skillful precipitation forecast (Fig. 2a,b). The COAMPS forecast likely does not 240

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capture the detailed structure of the observed precipitation at this resolution in part because finescale orographic features, such as the coastal ranges, are not well resolved. Despite these biases,
the analysis undertaken here are still valid as they are focused on understanding the physical
mechanisms that lead to changes in the forecast. Modest values of precipitation are evident in
Central California around Morro Bay and San Luis Obispo in both the Stage-IV and the CTL run
(Fig. 2a,b). As will be discussed in the next section, this is the region where the most substantial
precipitation enhancements occur in the perturbed run.

248

b) Initial Condition Perturbations

The COAMPS moist adjoint model provides an objective method for determining 249 sensitivities of the response function (18-h accumulated precipitation averaged within the box 250 shown in Fig. 2) to the initial condition state variables. Informally speaking, this means that the 251 252 output of the adjoint model produces the amount by which the 18-h accumulated precipitation will change for a given perturbation in a state variable made at the initial time. As described in 253 254 Section 2a, the perturbations that are employed provide the largest change to the 18-h accumulated precipitation possible within the limit of observational uncertainty and are thereby 255 referred to as the optimal perturbations. These optimal perturbations may be conceptualized as 256 patterns highlighting the regions where the forecast is most sensitive to changes in the initial 257 state. For clarity, the various simulations' acronyms used within this study are shown in Table 1. 258 The nonlinear control simulation is simply CTL, the perturbed tangent linear model's forecast is 259 *LPF* with difference from the control termed *LPF* – *CTL*, and lastly the perturbed nonlinear 260 simulation's forecast is NPF with difference from the CTL termed NPF – CTL. 261

The initial condition (0h) LPF – CTL are shown in Fig. 3 with the CTL run in color fill
and the LPF – CTL in the black contours such that solid contours overlaid on positive CTL

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264 values indicate an enhancement of the LPF values relative to the CTL. As seen in Fig. 3a,b, the 850-hPa optimal perturbations of wind speed and moisture content are situated within and 265 surrounding the tropical moisture plume, or AR, between 130°W and 140°W. The 850-hPa 266 moisture is enhanced by up to 0.4 g kg⁻¹ within a large fraction of the AR core, around 23°N, 267 140°W, and decreased on the south east edge, around 20°N, 130°W (Fig. 3b). The 850-hPa wind 268 speed optimal perturbations are seen to increase the wind speed on the tail of the AR (~23°N, 269 140°W) and near the south eastern edge (~20°N, 130°W); maximum values of these perturbations 270 are up to about 0.15 m s⁻¹. The optimal perturbations at the 500-hPa level are situated much 271 farther upstream of the AR compared to the lower levels, and are located near a jet entrance 272 region (Fig. 3c,d). The fact that the perturbations occur farther upstream with increasing altitude 273 is consistent with the advective speed being much greater at upper levels; if these perturbations 274 are to impact the forecast during the same time period as the lower levels, they must be situated 275 farther away. This is also consistent with the structure of the sensitivities being tilted against the 276 shear (Doyle et al. 2014, 2019; Reynolds et al. 2019) and can be interpreted as growing through 277 potential vorticity (PV) unshielding, which is the result of a superposition of PV maxima, as in 278 the Orr mechanism (Orr 1907; Badger and Hoskins, 2001; Reynolds et al. 2001). 279

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281 **4. Results**

a) Evolved Perturbations

The remainder of the manuscript will be focused on the evolution of the optimal
perturbations and the dynamical differences between the CTL and the LPF. The difference in 48h accumulated precipitation between CTL and LPF is shown in Fig. 2c and exhibits a long

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286 narrow band of enhanced precipitation centered at the latitude of Central California (~35°N) and a modest reduction in precipitation poleward of 36° N. The dipole structure of the precipitation 287 band suggests a shift in the LPF's AR position, however, as will be demonstrated later in this 288 section, the first order change is in the AR's strength. The fact that the band of enhanced 289 precipitation extends far offshore, away from any topographic features, indicates that it is driven 290 in part by a dynamical mechanism. Similarly, the enhancement over land may be suggestive of 291 orographic effects on precipitation seen at both the coastal mountain ranges of San Luis Obispo 292 (121W, 35N) and the southern Sierra Nevada mountains (119W, 35.5N). The optimal 293 294 perturbations are formulated to increase precipitation in the LPF compared to the CTL during the last 18 h of the 48-h forecast. The LPF minus CTL domain-averaged 18-h accumulated 295 precipitation is about 3 mm, which is evident when comparing the deeper blue colors to the 296 lighter brown in Fig. 2c. Though this may seem small, it is important to emphasize that this is the 297 overall regional average which is offset by some areas of reduced precipitation. 298

While the perturbations have been tailored to optimize the precipitation in a linear 299 context, the nonlinear growth of the perturbations may also be examined by making a nonlinear 300 301 forecast from the optimally perturbed initial state. It is important to compare the forecast of the 302 LPF with the NPF because substantial differences between the two runs may indicate the dominance of non-linearities, which are not represented within the LPF. The 48-h accumulated 303 precipitation of the NPF minus CTL, is shown in Fig. 2d. A comparison of the difference fields 304 305 (Fig. 2c,d) shows that they are qualitatively similar, indicating that within the 48-h forecast, the non-linearities are sufficiently small such that the linear and nonlinear precipitation perturbations 306 307 are similar. Therefore, we are confident in moving forward with analyzing the LPF in more 308 detail.

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309	In order to perform a diagnostic analysis on the differences between the CTL and LPF
310	runs, it is necessary to clarify when in time the evolved perturbations may be leading to the
311	precipitation enhancement. A time series is shown in Fig. 4 of the average accumulated
312	precipitation at 15-minute intervals in a region that isolates the precipitation enhancement
313	between the bottom of the response function box up to 36°N indicated by the bold, black dotted
314	line in Fig. 2c. The LPF's enhanced precipitation period begins at 26h, four hours prior to the
315	response function accumulation period. This is followed by strong growth in the precipitation
316	enhancement that occurs for about 10 h from 26h to 36h, after which time both the CTL and LPF
317	continue to accumulate precipitation at about the same rate.
318	The same method of averaging is applied to the IVT field to quantify the role of the
319	orographic precipitation forcing (i.e. moisture transport; Neiman et al. 2016; Martin et al. 2019)
320	within the LPF's enhanced precipitation field (Fig. 4). As can be seen here, elevated values of
321	the LPF's IVT begin at about 24h, just prior to the start of the period of enhanced precipitation,
322	and end at the same time (36h). Furthermore, it should be noted that while the percentage
323	increase in IVT in Fig. 4 may seem modest, a map of the difference in IVT between the LPF and
324	CTL (not shown) has a complex pattern of positive and negative values, suggesting that IVT can
325	be more locally enhanced than what is reflected in the regional average. Since moisture transport
326	has been shown to correlate well with orographic precipitation (Neiman et al. 2002, 2009a;
327	Ralph et al. 2013), this increase in IVT supports the hypothesis that the observed precipitation
328	enhancement is partially due to the increased orographic precipitation forcing.
329	To further investigate the forcing for orographic precipitation, we examine the
330	components of low-level water vapor transport within the 10-h time period of precipitation
331	enhancement in more detail. From 24h through 36h, a warm front propagates ahead of the AR

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(Fig. 5e-h). In this same area as the moisture plume, strong low-level wind speeds directed from 332 the SSW are evident in the CTL run (Fig. 5a–d), indicating a region of very large moisture 333 transport. At 24h (Fig. 5a,e), 2 h before the enhanced precipitation time period, a strengthening is 334 observed in both the 850-hPa LPF – CTL wind and moisture content behind the warm front. 335 Over the course of the next 8 h, through 32h, the LPF – CTL wind speed enhancement intensifies 336 337 substantially while simultaneously propagating north-eastward and making landfall (Figs. 5b,c). At 28h, the moisture enhancement intensifies while remaining coincident with the region of 338 enhanced wind speed (Fig. 5f). By 32h (Fig. 5g), the moisture is no longer enhanced behind the 339 340 warm front, likely a result of the greater precipitation taking place. Finally, at 36h (Fig. 5d,h) both the LPF – CTL wind and moisture content are reduced behind the warm front. As 341 mentioned previously, the increase in vapor transport supports the hypothesis that the 342 precipitation enhancement is partially due to a greater orographic precipitation forcing. 343

During this same time period, from 24h through 36h, an intensifying narrow band of 344 moisture convergence $(-q_m \nabla \cdot \vec{u})$ and low-level PV is situated behind the warm front in the CTL 345 346 (Fig. 6). At 24h (Fig. 6a,e), there is already substantially enhanced moisture convergence behind the warm front in the LPF - CTL, which results in diabatic generation of PV. Over the next 8 h, 347 348 both the moisture convergence and PV are enhanced in LPF – CTL while the system is making landfall (Fig. 6b–c,f–g). Since the CTL fields are also intensifying during this time period, the 349 fact that the LPF – CTL field is enhanced at these times shows that the rate of intensification for 350 these quantities is greater in the LPF than the CTL. By 36h, the coherent moisture convergence 351 dipole structure off the CA coast in the LPF – CTL is less apparent while the PV dipole appears 352 relatively unchanged (Fig. 6d,h). Just as Fig. 5 indicates that the forcing for orographic 353

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precipitation is enhanced within the LPF, Fig. 6 shows that dynamically driven precipitation isenhanced through moisture convergence.

356 To provide analysis of the vertical structure, Fig. 7 shows a cross section taken across the 357 warm front (location shown in Fig. 5c) at the start of the response function integration time (forecast hour 30) in which the precipitation enhancement is occurring. At this time, the CTL 358 exhibits strong wind speeds of up to 25 m s⁻¹ at 900 hPa situated on the southeast side of the low-359 360 level PV anomaly centered on the front (Fig. 7a,b). A dipole structure is observed in the LPF – CTL low-level wind speed (Fig. 7a), where enhancements of up to 6 m s⁻¹ indicate the clear 361 362 development of a low-level jet at 900 hPa which was not present in the CTL simulation. At first glance, this may seem to be a shift in the wind speed locations, but examination of the total LPF 363 wind fields (not shown) compared with that of the CTL clearly shows the winds are enhanced up 364 to 20% rather than shifted. In fact, the location of this dipole is centered on the PV enhancement 365 (Fig. 7b), which matches the expected horizontal circulation change due to low-level latent 366 heating as investigated by Lackmann (2002). Importantly, the mixing ratio enhancements (Fig. 367 7d) are found to overlap the region of strengthened wind speed, consistent with the observed 368 369 enhancement of IVT. The reason that the region of enhanced vapor transport is not directly 370 coincident with the enhanced precipitation in this cross section is because the section location is well offshore where orographic influences are not present. 371

Strong vertical motion that is vertically sloped and oriented along equivalent potential surfaces occurs in the CTL (not shown) and also approximately along the region of moisture convergence (Fig. 7c,e). Examination of equivalent potential temperature and absolute momentum (not shown) clearly indicate that this sloped region is symmetrically unstable, demonstrating that perturbations are expected to grow rapidly there. Contrastingly, the LPF –

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377 CTL vertical motion clearly has two distinct regions, one with a slantwise orientation, similar to that of the CTL, and the other with a distinct vertical orientation (Fig. 7c,e) consistent with deep 378 convection as in Cordeira et al. (2013). Cross sections of convective available potential energy 379 (CAPE; not shown) indicate that the LPF has an environment that is more conducive to 380 convection. A greater degree of convection is likely also a component of the precipitation 381 382 enhancement, although this is not investigated further here, and thus remains an open question for future work. As demonstrated by the clear increase in both the vertical ascent and vapor 383 384 transport, the cross sections further support the hypothesis that the observed precipitation 385 enhancement is a result of contributions from the dynamic precipitation component and orographic precipitation forcing. 386

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b) The Transverse Circulation

The Sawyer-Eliassen transverse circulation model, which produces the vertical and cross-388 frontal ageostrophic circulation perpendicular to the upper level jet structure in an approximately 389 390 straight flow, is used to further investigate the dynamical component of the precipitation enhancement by separating the vertical velocity and moisture convergence into moist and dry 391 components. Here, the moist processes are due to the latent heating effects while the dry forcing 392 components arise from quasi-geostrophic shearing and stretching deformation contributions (see 393 Keyser and Shapiro 1986 and Winters and Martin 2014 for a more in-depth description). The dry 394 395 and moist components are obtained by solving the equation in a piecewise manner by using the dry and moist forcings individually within the PDE solver. Note that Figs. 8–9 show only a 396 subset of the full PDE solver's domain to match the previous cross sections in Fig. 7. 397

398 While the Sawyer-Eliassen model directly yields the components of the transverse 399 circulation (v_{ag}, ω) , we investigate the moisture convergence rather than v_{ag} because it is more 18

400 relevant for explaining the precipitation enhancement. The Sawyer-Eliassen derived moisture convergence is calculated in three steps: 1) solve the Sawyer-Eliassen equation for the 401 ageostrophic stream function (Ψ), 2) calculate the cross-jet ageostrophic wind using the equation 402 $v_{ag} = -\frac{\partial \psi}{\partial p}$, and 3) assume 2D convergence in the plane of the cross section such that 403 $MConv = -q_m \frac{\partial v_{ag}}{\partial y}$ where y is the cross-jet streak direction. The Sawyer-Eliassen derived total 404 405 moisture convergence (Fig. 8a) is calculated along the same transect as Fig. 7, and compares very well with the full 3D model calculated moisture convergence in regard to strength, slope, 406 407 and vertical extent (Fig. 7e). Similarly, the Sawyer-Eliassen model reproduces the LPF - CTL 408 moisture convergence very closely in terms of strength, slope, and vertical extent (solid black 409 contours), though it has a maximum at the surface (Fig. 8a), while the real case (Fig. 7e) is slightly elevated from the surface. Therefore, despite the assumptions included in the Sawyer-410 411 Eliassen equation, it still provides a very accurate depiction of the transverse circulation in this 412 case.

413 The main utility of the Sawyer-Eliassen model for this analysis is the ability to separate 414 the transverse circulation into the dry and moist components. The geostrophic (dry) and the 415 diabatic (moist) components of moisture convergence shown in Fig. 8b and Fig. 8c, respectively, 416 clearly illustrate that the moist component is dominant. The dry component is constrained to the very lowest levels (Fig. 8b) with peak values being much smaller than those observed (Fig. 7e). 417 Conversely, the moist component has peak values, slope, and vertical extent that are much closer 418 to that of the CTL (Fig. 8c). With that said, the dry component is not negligible. Furthermore, it 419 420 is important to be cognizant that this analysis is only at a single time, which raises the question of what role the dry component had at an earlier time. Did the dry component have much greater 421

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strength earlier on? Did the LPF – CTL dry component at an earlier time initiate the LPF – CTL
moist component? These questions will be addressed in Section 4c.

424 Vertical velocity can also be calculated from the Sawyer-Eliassen analysis by using $\omega =$ $\frac{\partial \psi}{\partial v}$ (Fig. 9). Comparison of the total Sawyer-Eliassen derived vertical velocity (Fig. 9a) to that of 425 the 3D model output (Fig. 7c) indicates it is also accurate, though not to the degree of the 426 moisture convergence comparisons. The CTL vertical velocity derived by the Sawyer-Eliassen 427 428 method retains the sloped orientation and vertical extent but is biased low in magnitude. This is attributed to the fact that it is a modestly convective region as indicated by the CAPE values 429 (~300 J kg⁻¹; not shown), which is not accounted for by the Sawyer-Eliassen analysis. Similarly, 430 431 while the slope and vertical extent of LPF perturbation vertical velocities are reproduced well, there is a low bias in magnitude. It is important to note that the vertically upright orientation of 432 LPF perturbations (Fig. 9a) is reproduced very well in the Sawyer-Eliassen model implying that 433 434 it is not only due to convection but also a result of the forced ascent. As in the results for 435 moisture convergence, the vertical velocity is dominated by the moist component term at this 436 time.

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c) Pseudo-Lagrangian Evolution of the Transverse Circulation

The transverse circulation analysis above applies only to one cross section at a single time, so a complimentary analysis is undertaken by initializing backward trajectories from the central point along the cross section of Figs. 7–9 to investigate a system-relative evolution of the transverse circulation. At each point along each trajectory in Fig. 10a, a cross section is drawn to be nearly perpendicular to the upper-level jet streak (not shown). This establishes a pseudo-Lagrangian cross section time series for each trajectory. It is important to note that it is called

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"pseudo-Lagrangian" because it assumes that the entire cross section moves together, which, of
course, is not the case. Despite this caveat, animations of the cross sections along the pseudoLagrangian trajectory produce what seem to be an accurate depiction of how the system-relative
cross-section evolves (not shown).

While the pseudo-Lagrangian cross section time series can address whether the results in 448 449 Section 4b are applicable for other times, it does not address the spatial issue. A simple way to investigate how the results from Section 4b may vary spatially is to initiate the backward 450 trajectory at the same point as in Fig. 10a, but at different times. Since the system is moving, this 451 452 will ensure that different portions of the AR are sampled. This was done for backward trajectories initiated at 24h through 36h with each air parcel moving all the way backwards to the 453 454 initial time at 0h (Fig. 10a). This method produces 13 pseudo-Lagrangian cross section time series each for CTL and LPF from which the following analysis is comprised. 455

The strength of the Sawyer-Eliassen transverse circulation for each component is 456 457 measured from the maximum value of $-\omega$ for each time step along the pseudo-Lagrangian cross section. This is repeated for each of the 13 pseudo-Lagrangian trajectories, and then averaged by 458 aligning them on the respective final trajectory times (Fig. 10b). Consistent with previous results, 459 the strength of the LPF – CTL vertical velocity is found to be substantially greater than that of 460 the CTL. Furthermore, the moist component is dominant from about 10h onwards in both the 461 CTL and LPF; the dry component remains very steady from beginning to end. This analysis 462 shows that dominance of the moist component of the transverse circulation applies not only to a 463 single transect at a single time, but consistently influences the dynamics of the transverse 464 465 circulation for the system as a whole.

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A major benefit of creating a time series of the transverse circulation strength is that it 466 allows us to form conclusions about its evolution. It reveals that the LPF's vertical velocity 467 enhancement is initiated from the moist component starting at about 12h (Fig. 10b), at which 468 time the slope of the LPF curve becomes greater than the CTL. Sustained LPF vertical velocity 469 470 growth continues up until about 26h at which point the CTL remains steady, and the LPF begins 471 to decline in strength. The CTL and LPF total vertical velocities are clearly controlled by their respective moist components with the dry components providing a small, but steady contribution. 472 Both the CTL and LPF have very similar dry component strengths, indicating that the quasi-473 474 geostrophic shearing and stretching environments are approximately the same between the different runs. 475

The decomposition of the moist and dry forcings in the Sawyer-Eliassen transverse circulation analysis suggests the possibility of a positive feedback mechanism. Latent heating at the developing frontal boundary provides a positive forcing to the Sawyer-Eliassen circulation. This positive forcing in turn leads to moisture convergence, which may then drive further latent heat release. While it is outside of the scope of this study to prove that this is indeed a feedback mechanism, it is still clear that the dynamical component of the vertical velocity is highly dependent on the nature of the moist processes and, thereby, on the moisture content.

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d) Lagrangian Evolution of the Optimal Perturbations

484 Until this point, this study has been focused on the evolved perturbations with only some 485 minimal connections to the initial optimal perturbations. To address this gap, forward trajectories 486 are utilized to investigate how the optimally perturbed regions may connect to the downstream 487 processes described previously. Forward trajectories are initialized on the 700-hPa level, where 488 the largest moisture perturbations were made, at 0h in a 11x11 square block with a 40-km

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spacing and centered on the maximum positive initial condition moisture perturbation (Fig. 11a).
This will enable the initial perturbations to be connected to the moist transverse circulation
analysis already discussed. It is clear from Fig. 11a that the LPF trajectories are more tightly
clustered than the CTL trajectories, consistent with stronger convergence from the enhanced
transverse circulation.

494 The average Lagrangian (along air parcel trajectory) moisture content, PV, and pressure are shown in Figure 11b for the CTL and LPF. To ensure that the shape of the averaged curves is 495 representative of the individuals, each trajectory is shifted in time to match a clear rise in 496 497 moisture content (Fig. 11b, at 23h) immediately followed by a sharp decline, which is a feature found to be present among all trajectories. The average time axis among all of the trajectories is 498 499 used in order to provide a sense of the time scale. The three subplots in Fig. 11b illustrate a 500 three-phase sequence, which is not only very consistent with the previous results pertaining to the transverse circulation, but also ties the sequence of events back to the initial perturbations. At 501 the start of phase I (6h to 20h), it is clear that the LPF air parcels have enhanced moisture 502 content, a direct result of the perturbation. As phase I progresses, both the CTL and the LPF 503 moisture content have a very small, but steady increase, with the LPF maintaining greater 504 505 moisture throughout. No substantial changes occur for PV or pressure during this phase for these trajectories. 506

507 Phase II (20h to 24h) is short but distinctly indicates that the growth rate of the moisture 508 content in both the LPF and the CTL increases noticeably. During this time, a rise in PV 509 indicates that diabatic generation has begun, with the perturbed LPF exhibiting a substantially 510 greater increase than CTL. Both observations are consistent with the moisture convergence 511 processes previously described, which can cause an increase in moisture content and serve to

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512 diabatically generate PV. Throughout phase II, the air parcel pressure remains mostly steady513 with only a modest decrease in height (increase in pressure).

514 Lastly, in phase III (24h to 30h), the air parcels begin their rapid ascent in the warm 515 conveyor belt (Wernli and Davies 1997; Wernli 1997; Schemm et al 2013) which is also confirmed from the plan-view sequence shown in supplemental Fig. 1. The PV reaches a 516 517 maximum, and the moisture content rapidly decreases, likely due to precipitation. During the first few hours of phase III, the LPF ascent rate is initially much greater than that of the CTL 518 519 and, in turn, it reaches a higher overall elevation by 30h. The LPF's enhanced ascent rate and PV 520 is consistent with the analysis of the transverse circulation in Sections 4b-c in that stronger moist processes lead to greater ascent. This coherence suggests that the physical process leading to the 521 522 observed Lagrangian evolution in Fig. 11b are actually those described by the transverse circulation. 523

A similar analysis is performed by initializing another 11x11 box of air parcels centered 524 525 on the negative moisture perturbation at approximately 26.5°5N, 135.5°5W. The parcel trajectories (not shown) terminate in Southern CA and Northern Baja, Mexico (117W, 32N), 526 which is outside of the response function domain. Therefore, the Lagrangian approach to 527 understanding the mechanism for increasing precipitation is not applicable to this region, and the 528 impact on precipitation from these perturbations must be coming from changes to the evolving 529 530 dynamics not captured in a Lagrangian framework. For example, it may be that the negative perturbations are associated with changes to the initial cyclone (AR wave 1) and not the frontal 531 wave (AR wave 2). 532

The consistency between the Lagrangian evolution of the optimal moisture perturbation
 with the enhancements observed in the transverse circulation provides a connection between the
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535 initial perturbations and their downstream modifications of important physical processes. The forward trajectory analysis supports the hypothesis that the positive moist optimal perturbations 536 are positioned, at least in part, to provide more moisture available for enhanced latent heat 537 release. Enhanced latent heating leads to an increase in the dynamical component of the 538 precipitation through the enhanced transverse circulation. This analysis demonstrates the utility 539 540 of the Lagrangian air parcel trajectory method in tracking the evolution of the optimal perturbations. While outside the scope of this work, a future study could initialize air parcels in 541 this way for different areas within all the largest initial condition perturbation areas and track 542 543 their evolution.

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545 **5. Conclusions and Discussion**

This study is focused on understanding the nature of rapid perturbation growth in a 546 tangent linear model for an extreme landfalling AR. The perturbed tangent linear forecast (LPF) 547 has been optimized following Doyle et al. (2019) to enhance the final 18h of the accumulated 548 precipitation in an area of California and within the constraints of analysis uncertainty. In 549 agreement with Cannon et al. (2018), the analysis presented here demonstrates that the 550 precipitation enhancement in the perturbed run is a result of both the enhanced orographic 551 precipitation forcing and the dynamic precipitation component. Importantly, evidence suggests 552 553 that the enhancement of both components originate from the same latent heating process, thereby making the precipitation forecast particularly sensitive to errors in initial condition moisture 554 555 content.

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556 The sequence of events in which the moisture perturbations enhance precipitation is illustrated schematically in Fig. 12. At T+00, the positive moisture perturbation is initiated on the 557 periphery of the AR where it is gradually fed into the developing frontal zone. Starting at T+12, 558 PV is diabatically generated at the frontal zone through moisture convergence and the 559 560 subsequent release of latent heat. The wind field responds to the PV anomaly by 561 increasing/decreasing on the warm/cool side of the front following Lackmann (2002), thereby leading to a local enhancement of moisture transport. The transverse circulation responds to the 562 latent heat release by strengthening the moisture convergence and ascent at the frontal zone. 563 564 During this development, the moisture perturbation remains mostly unchanged and continues its way toward the front. Finally, at T+24, the moisture perturbation reaches the frontal zone leading 565 to substantial latent heat enhancements. This additional heating enhancement, along with the 566 sustained latent heating independent of the perturbation, drives an even greater moisture 567 transport through the generation of a stronger PV anomaly, and an amplification in the transverse 568 569 circulation as described by the Sawyer-Eliassen model. The observation that the initial moisture perturbation is fed into the region of moisture 570

convergence aligns well with the results of Reynolds et al. (2019) and Doyle et al. (2014, 2019).
They found that the optimal perturbations within ARs exhibit upshear tilts which can extract
energy from the mean flow through the PV unshielding mechanism. The results presented here
are consistent with the hypothesis that the physical "reason" why the moist adjoint sensitivity
adds moisture within the AR for this case is so that it may serve to enhance latent heating
processes and thereby enhance both the orographic and dynamic components of the precipitation.

577 Clearly, moist processes are critically important in the development of the strong
578 transverse circulation, which is itself responsible for the strong moisture convergence and latent

579 heating. Since the Sawyer-Eliassen model is also forced by a latent heating term, these results suggest a feedback mechanism whereby the latent heating is both a source and a consequence of 580 the transverse circulation. This may be important from a predictability standpoint because it 581 demonstrates a mechanism by which small changes in the initial moisture content can lead to 582 large changes in the forecast downstream. For example, targeted observational campaigns should 583 584 closely consider sampling in the moist regions which feed into the transverse circulation since any error in those regions can grow substantially. Furthermore, since the transverse circulation is 585 frequent component of ECs, and ECs are often accompanied by ARs (Zhang et al. 2018), the 586 587 results of this case study are likely to be representative of a large number of ARs.

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Figures

Acronym	Description
CTL	Control nonlinear forecast
LPF	Linear Perturbed Forecast: The sum of the control nonlinear forecast and the optimal perturbation evolved using the Tangent Linear Model.
NPF	Nonlinear Perturbed Forecast: The nonlinear forecast run from an analysis which is the control analysis plus the initial optimal perturbation.
LPF – CTL	Linear Perturbation: The optimal perturbation evolved using the Tangent Linear Model, equivalent to the difference between CTL and LPF.
NPF – CTL	Nonlinear Perturbation: The difference between CTL and NPF.

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Table 1: The descriptions of acronyms used in this study.



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Figure 1: Plan-view maps every 12 hours of the a)–e) control run IVT (kg s-1 m-1; shaded) and 925-hPa potential temperature (contour interval of 2 K; solid gray), and f)–j) 250-hPa wind speed (m s-1) and SLP (contour interval of 4 hPa; solid black). The surface fronts are shown in (a – d) with the cold front in black dashed and the warm front in solid black. The location of a transect for Figures 7–9 is shown in (i) as a solid black line. The "W1" and "W2" labels refers to different AR waves, and the "C" label refers to a developing secondary cyclone (see text for discussion).



48-h Accumulated Precipitation

Figure 2: Plot of the a) Stage-IV and b) CTL 48-h accumulated precipitation (mm), and c) linear perturbations (LPF minus CTL) and d) nonlinear perturbations (NPF minus CTL) 48-h accumulated precipitation difference fields. The response function box is shown in the solid black box. The thick horizontal dotted black line across the response function box in c) separates the precipitation enhancement and reduction regions used in Figure 4 within the box. The precipitation enhancement is identified in c).



Figure 3: The initial condition perturbations for the a) 850-hPa wind speed with IVT contours (kg s-1 m-1; shaded), b) 850-hPa mixing ratio, (g kg-1; shaded), c) 500-hPa wind speed (m s-1; shaded), and d) 500-PV with the LPF-CTL in solid black (positive) and dashed black (negative) contours for each respective field. The solid gray contours in each plot are the equivalent potential temperatures on the respective levels with contour interval (ci) = 4 K and five passes of a 5-point local smoother applied.



Figure 4: Time series of average accumulated precipitation (black) and the IVT (blue) within the sub-region of Figure 2c (see text). The control (CTL) run is shown in the solid lines and the linearly perturbed (LPF) run in dashed.



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Figure 5: Plan view every 4 hours beginning from 24 h of the 850-hPa a)–d) wind speed (m s-1; shaded), and e)–h) water vapor mixing ratio (g kg-1; shaded). The LPF-CTL fields are shown in black with solid being positive and dashed being negative values. The approximate position of the low-pressure center of the developing secondary cyclone is labelled "L" and the warm front indicated by the thick black curved line. The position of a transect used in Figs. 7–9 is shown in c) in the thin black straight line.



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Figure 6: Plan view every 4 hours beginning from 24 h of the 850-hPa a)–d) PV (PVU; shaded), and e)–h) moisture convergence (10-3 g kg-1 s-1; shaded; positive is converging). The LPF-CTL fields are shown in black with solid being positive and dashed being negative values. The approximate position of the low-pressure center of the developing secondary cyclone is labelled "L" and the warm front indicated by the thick black curved line. The position of a transect used in Figs. 7–9 is shown in c) in the thin black straight line.



Figure 7: Cross sections at 30 h along the transect shown in Fig. 5c with "NW" being the northwest side and "SE" being the southeast side of a) wind speed (m s-1; shaded), b) PV (PVU; shaded), c) negative omega (Pa s-1; shaded), d) mixing ratio (g kg-1; shaded), and e) moisture convergence (10-3 g kg-1 s-1; shaded). The LPF-CTL fields are shown in black with solid being positive and dashed being negative values. The sub plot below each transect shows the accumulated precipitation for the CTL (solid black) and the LPF (dashed blue).



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Figure 8: Cross sections at 30 h along the transect shown in Fig. 1i of the Sawyer-Eliassen derived moisture convergence (10-3 g kg-1 s-1; shaded) for the a) total, b) quasi-geostrophic (dry) component, and c) diabatic (moist) component. The LPF-CTL fields are shown in black with solid being positive and dashed being negative values.



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Figure 9: Cross sections at 30 h along the transect shown in Fig. 1i of the Sawyer-Eliassen derived vertical velocity (10-3 g kg-1 s-1; shaded) for the a) total, b) quasi-geostrophic (dry) component, and c) diabatic (moist) component. The LPF-CTL fields are shown in black with solid being positive and dashed being negative values.



Figure 10: a) Plan-view of the initial time IVT (kg s-1 m-1; color fill) and SLP (contoured every 4 hPa; solid gray) with the backward trajectories of the CTL (black) and LPF (blue). The hourly locations of the air parcels are shown in the filled circle along each trajectory. The final time

transect is shown as a reference in the solid black line. The trajectories are initialized at 700 hPa and are separated in time at one hour intervals starting from 24 h to 36 h. b) The average Sawyer-Eliassen - ω (hPa s-1) along the trajectories shown in a). The total value of - ω is in black with the moist component in red and the dry component in blue. The CTL is shown in the solid lines with the LPF in the dashed.



Figure 11: a) A plan-view of the initial time IVT (kg s-1 m-1; color fill), SLP (contoured every 4 hPa; solid gray), and initial condition LPF-CTL (black contours with solid being positive and dashed being negative values). Forward trajectories are shown for CTL (black) and LPF (blue)

with the initialization points shown in black filled circles. b) From top to bottom: the mixing ratio (g kg-1), PV (PVU), and pressure (hPa) following the air parcel trajectories shown in a) for the CTL (solid, black) and LPF (dashed, blue). The average of all trajectories is shown in bold with one standard deviation from the mean shown in the transparent color fill. The trajectories in b) are shifted in time to center each trajectory on the same feature (described in text).



Figure 12: A pseudo-4D schematic illustrating the evolution of cross sections in an AR relative reference frame. The IVT values of the AR are shown in the yellow to red shading (small to large values respectively). The cross sections show potential temperature (gray contours), the LPF moisture perturbation (black contours), and the transverse circulation in plane of the cross section in the black arrows with the strength indicated by the length and arrow thickness.