Role of Moist Processes in the Tracks of Idealized Midlatitude Surface Cyclones

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ABSTRACT

The effects of moist processes on the tracks of midlatitude surface cyclones are studied by performing idealized mesoscale simulations. In each simulation, a finite-amplitude surface cyclone is initialized on the warm-air side of a zonal baroclinic jet. For some simulations, an upper-level cyclonic anomaly upstream of the surface cyclone is also added initially. Sensitivities to the upper-level perturbation and moist processes are analyzed by both performing a relative vorticity budget analysis and adopting a potential vorticity (PV) perspective.

Whatever the simulation, there is a systematic crossing of the zonal jet by the surface cyclone occurring after roughly 30 h. A PV inversion tool shows that it is the nonlinear advection of the surface cyclone by the upper-level PV dipole, which explains the cross-jet motion of the surface cyclone. The simulation with an initial upper-level cyclonic anomaly creates a stronger surface cyclone and a more intense upper-level PV dipole than the simulation without it. It results in faster northward and slower eastward motions of the surface cyclone.

A moist run including full microphysics has a more intense surface cyclone and induces a faster northward motion than the dry run. The faster eastward motion is due to the diabatically produced cyclonic circulation at low levels. The faster northward motion is explained by the stronger upper-level anticyclone due to released latent heat, together with the closer location of the surface cyclone to the upper-level anticyclone. Finally, a moist run with only condensation and evaporation exhibits less latent heat release and a slower northeastward motion of the surface cyclone than the full moist run.

1. Introduction

During recent decades, the forecast skill for high-impact weather events have been significantly improved (Thorpe 2004). The major short-range-forecast failures in the intensity and track of extratropical cyclones that occurred in the 1980s and early 1990s (e.g., Sanders and Gyakum 1980; Anthes et al. 1983; Reed et al. 1993a) would probably not occur nowadays with the current operational weather prediction models. Despite these improvements, the exact location of peak wind gusts at a lead time of 1–2 days is still difficult to forecast (Hewson et al. 2014). Furthermore, medium- to extended-range forecasts present synoptic-scale errors that may arise from the misrepresentation of the diabatic processes in the ascending air masses of extratropical cyclones and the so-called warm conveyor belts (Grams et al. 2011; Davies and Didone 2013). Motivated by these forecast challenges, there is a recent regain of interest in the role played by cloud microphysics in producing PV anomalies within extratropical cyclones and, in particular, warm conveyor belts (Joos and Wernli 2012; Schemm et al. 2013; Chagnon et al. 2013; Martinez-Alvarado et al. 2014). The present paper goes in the same vein and aims to show how diabatically produced PV anomalies within the warm conveyor belts affect the tracks of extratropical cyclones.
There is an extensive literature on the effect of moist processes in the intensity of extratropical cyclones (e.g., Uccellini 1990; Chang et al. 2002). Most studies show a more rapid cyclone deepening in presence of moist processes because of condensational heating in regions of upward motion. This has been shown by theoretical studies on moist baroclinic instability (Mak 1982; Emanuel et al. 1987), idealized numerical studies (Gall 1976; Lapeyre and Held 2004), numerical studies of real storms (Kuo and Reed 1988; Reed et al. 1993a,b; Davis et al. 1993), and diagnostic studies based on observational datasets (Danard 1964; Smith et al. 1984). As moisture content rises and latent heat is released, the storm strength increases in terms of central pressure minimum, winds, and precipitation, as revealed in idealized modeling (Booth et al. 2013). Despite this general agreement between existing theories, idealized modeling, and more realistic studies, the amount of cyclone deepening due to moist processes is still debated. For instance, no quantitative relation can be obtained from moist baroclinic instability to interpret realistic scenarios. It is strongly dependent on the parameterized effects of moisture, the assumed background humidity profile, and the degree of nonlinearities (de Vries et al. 2010).

One simple way of understanding the intensification of the cyclone by latent heat release is through vertical velocity. The vorticity is intensified by an enhanced upward motion in the warm sector of the cyclones because of the released latent heat. The increased vertical velocity can be explained by the fact that condensational heating creates an additional forcing term in the omega equation (Danard 1964; Smith et al. 1984) and by the presence of saturation, which reduces the static stability and tends to amplify the response to a given forcing in the same omega equation (Reed et al. 1993b).

In terms of energetics, condensational heating within extratropical cyclones creates a source of eddy available potential energy but also tends to reinforce the transfer of energy from the mean-flow available potential energy to the eddy available potential energy and from the eddy available potential energy to eddy kinetic energy (Danard 1964; Robertson and Smith 1983; Chang et al. 1984). However, note that long-term moist simulations (warmer climate simulations) do not necessarily lead to more intense extratropical cyclones compared to dry simulations (present climate simulations), as condensational heating may have the reverse effect, on average, as it tends to reduce the zonal-mean available potential energy (Pavan et al. 1999; Lainé et al. 2011).

When considering potential vorticity (PV), latent heating creates positive and negative PV anomalies in the lower and upper troposphere, respectively (Hoskins et al. 1985; Reed et al. 1992; Stoelinga 1996). This happens in the region of maximum ascent, which occurs over and ahead of the cyclone. The positive low-level PV anomaly reinforces the surface cyclone circulation, which is already strengthening as a result of baroclinic interaction (Davis et al. 1993). At the mature stage of the cyclones, there is no evidence that the effect of latent heat release on the development of extratropical cyclones is a mechanism that is independent of the dry baroclinic instability/interaction. It, rather, acts more as an amplifier of the baroclinic instability/interaction (Davis et al. 1993; Langland et al. 1996; Rivière et al. 2010). This amplification varies from case to case, from one- to two-thirds of the cyclone deepening (Grönås 1995; Fink et al. 2012). Though the proportion of the impact of diabatic processes on the development of cyclogenesis varies a lot, the amplification due to latent heat release always happens, regardless of the strength of the cyclone. Zimmerman et al. (1989) showed it for a weak extratropical cyclone and Wernli et al. (2002) for the extreme winter storm Lothar.

While the influence of moist processes in the intensification of extratropical cyclones has been the subject of numerous studies, only a few publications mentioned the effect of latent heat release on their tracks. Idealized baroclinic instability studies show that the eastward phase speed increases with the heating intensity (Mak 1982). Simulations of real storms usually confirm the previous theoretical result by showing that the translational speed of the cyclone, which usually moves northeastward, is faster in moist simulations than in dry simulations (Davis et al. 1993; Grönås 1995; Rivière et al. 2010). The distance between the low centers in the moist and dry runs at the lead time of 1 day varies from almost 100 km to several hundred kilometers, depending on the case study (e.g., Rivière et al. 2010; compare Figs. 2 and 5). An easy way of understanding this effect is to consider the lower-level PV tendencies created by latent heat release. The diabatically produced PV anomaly below the region of maximum ascent not only strengthens the already existing vorticity but also accelerates its eastward propagation, as is well-known in the case of a diabatic Rossby wave (Parker and Thorpe 1995). However, the upper-level PV created by latent heat release acts to decelerate the eastward displacement by propagating westward (de Vries et al. 2010), so it is the importance of the lower-level component that explains the eastward motion. It is worth noting that the previously mentioned studies were not specifically focused on the effect of latent heat release on the motion speed of extratropical cyclones, and it is the main objective of the present paper to systematically address this issue. A particular attention will be paid to the motion of jet-crossing cyclones, which are
cycloons crossing the large-scale jet from its warm-air to cold-air sides. These cyclones are of particular interest, as they usually undergo a rapid deepening during and after the crossing of the jet axis (Wernli et al. 2002; Rivièr e and Joly 2006a,b; Fink et al. 2009). So, it is not so much the cyclone motion along the baroclinic zone that is the focus of the paper but, rather, the motion across the baroclinic zone.

A new mechanism has been recently proposed to explain the cross-jet motion of many extratropical cyclones and, more generally, the more northeastward-oriented cyclone tracks relative to the climatological mean flow (see, e.g., Palmén and Newton 1969, 95–97; Wallace et al. 1988). It corresponds to the so-called beta-drift mechanism generalized to the midlatitude atmospheric context. The recent results are based on quasigeostrophic numerical studies considering both zonally homogeneous (Gilet et al. 2009) and zonally inhomogeneous barotropic and baroclinic flows (Rivièr e 2008; Oruba et al. 2012, 2013). It is the positive vertically averaged background PV gradient due to the presence of the meridionally confined westerly jets that is the main driver of the cross-jet motion of the cyclones. This has been confirmed by the real case study of Rivièr e et al. (2012) using an operational numerical weather prediction model. When an upper-level trough interacts with a surface cyclone as in type-B cyclogenesis (Petterssen and Smeybe 1971), the positive background PV gradient is responsible for downstream dispersion of energy at upper levels and the building up of the downstream ridge. The zonally oriented upper-level dipolar PV anomaly made up of the upstream trough and downstream ridge tends to displace the surface cyclone poleward. As latent heat release is responsible for the reinforcement of the ridge aloft, we expect a faster cross-jet motion of surface cyclones in moist runs compared to dry ones, as suggested by Oruba et al. (2013). The paper is dedicated to check the validity of such a hypothesis by performing short-term dry and moist simulations using the Méso-NH mesoscale numerical model (Lafore et al. 1998). The simulations are composed of isolated synoptic-scale anomalies initialized south of a background westerly jet to analyze the processes involved in the cross-jet motion.

The article is organized as follows. Section 2 describes the model setup, the numerical experiments, and the diagnostic tools. Section 3 is dedicated to the results themselves. Relative vorticity and PV budgets are made to identify the various dynamical and physical processes influencing the cyclone motion. A PV inversion tool is also applied to attribute the displacement of the cyclones to different PV anomalies. Finally, concluding remarks and a short discussion are provided in section 4.

2. Methods

a. Model

The nonhydrostatic mesoscale research model Méso-NH (Lafore et al. 1998) is used. It is an Eulerian model with a gridpoint discretization on a staggered Arakawa C grid with a conformal projection system of horizontal coordinates and a Gal-Chen and Somerv ille (1975) system of vertical coordinates. The governing equations used in this study are based on the pseudo-incompressible system of Durran (1989), which is an anelastic approximation. A fourth-order centered advection scheme is used for the momentum components and the piecewise parabolic method (PPM) advection scheme (Colella and Woodward 1984) for other variables, associated with the leapfrog time stepping. A fourth-order explicit horizontal diffusion is applied only on the momentum components to suppress very short wavelength modes. With this dynamical configuration, Méso-NH has an excellent effective resolution [i.e., about (5–6)\(\Delta_x\), as shown in Ricard et al. (2013)]. Subgrid-scale vertical turbulent diffusion is described by a prognostic TKE-based scheme with a 1.5-order closure (Cuxart et al. 2000) with the turbulent length scale of Bougeault and Lacarrère (1989). The Coriolis force is considered. No radiation scheme and no surface scheme are used; free-slip conditions are specified at the lower boundary.

The dry simulation does not use any microphysical scheme and uses no water vapor. There are no diabatic effects for this simulation. The moist simulations use different microphysical schemes. The complete one is a one-moment bulk-cloud microphysics scheme with five prognostic hydrometeors (cloud droplets, rain, ice crystals, snow and graupel mixing ratios, and water vapor), which corresponds to the three-class ice parameterization (ICE3) scheme of Pinty and Jabouille (1998). This is called the “moist simulation.” The other simulation including moisture uses a simplified scheme that considers only the cloud water in addition to the water vapor. Only condensation and evaporation processes are in play, and cloud water is created if saturation is achieved. There is no other microphysical process in this simulation, no rain is generated, and it is called the “cloud water simulation.” This means that, at temperatures below the freezing point, cloud water content does not change, and, in ascending regions, cloud water accumulates. This corresponds to the closest moist simulation to the purely dry simulation and allows us to investigate the latent heat release associated with resolved-scale condensation only (Schemm et al. 2013).

b. Simulations setup

The model domain is 10 000 km long in the west–east direction and 6000 km wide in the south–north direction.
(500 × 300 points). The vertical extension is 16 km. The spatial resolutions of the model are Δx = Δy = 20 km and Δz = 500 m. The zonal boundary conditions are cyclic, and wall lateral boundary conditions (zero normal velocity) apply to the southern and northern boundaries.

The simulations are on an f plane centered at 40°N. The basic state of the simulations consists of a baroclinic zonal jet centered in the middle of the domain (Fig. 1). The basic-state zonal wind varies as a Gaussian function in both the horizontal and vertical directions. The maximum wind is located at the altitude of 8 km. The vertical profile of the Brunt–Väisälä frequency is fixed. In the troposphere, its mean value is 0.012 s−1 and varies from 0.01 to 0.014 s−1. It rapidly increases at 10 km high to reach 0.027 s−1 in the stratosphere. The basic-state temperature field is then calculated using the thermal wind balance. Finally, in the moist simulations, the initial relative humidity is set to 60% from 0 to 6.5 km high and 0% above. Some sensitivity to the initial humidity profile has been made, but it did not change the main findings of the paper.

Two types of perturbations are initially added to the basic state. The first one is composed of only one perturbation located near the surface and reaching a peak at the altitude of 1 km. Its latitudinal position is 1000 km south of the jet. The second one is composed of the same near-surface cyclonic anomaly as before added to a cyclonic anomaly reaching its maximum at 7.5 km near the tropopause and located to the northwest of the near-surface anomaly. This second configuration is shown in Fig. 2. The northwest position of the upper-level cyclonic anomaly relative to the surface cyclone corresponds to a typical synoptic configuration of explosive cyclogenesis (Sanders 1986). The first and second type of initial perturbation will be referred to as the one-perturbation and two-perturbations simulations, respectively. The perturbation streamfunction is expressed as follows:

$$\psi(x, y, z) = A \times \exp \left[ -\frac{(x-x_0)^2}{L} - \frac{(y-y_0)^2}{L} - \frac{(z-z_0)^2}{H} \right],$$

where A is the amplitude of the perturbation A = −5 × 10⁻⁶ m² s⁻¹; x₀, y₀, and z₀ correspond to the localization of the perturbation, L is the horizontal extent of the perturbation L = 7.5 × 10⁵ m, and H is the vertical extent of the perturbation, which varies from H = 4 × 10⁵, 5 × 10⁵, and 3 × 10⁵ m for the surface perturbation, the lower part of the altitude perturbation, and its upper part, respectively. Temperature is also modified so that it obeys the thermal wind balance.

The model is integrated up to 60 h, and the time step is 30 s. Multiple simulations are presented within this study, of which four will serve as the foundation of the paper: the one-perturbation and two-perturbations simulations in dry and moist configurations.
3. Results

a. Sensitivity of surface cyclone tracks to upper-level disturbances and moisture

Figure 3 compares moist and dry simulations with and without the initial upper-level disturbance at \( t = 48 \) h. The addition of an upper-level perturbation causes the surface cyclone to be stronger and located more to the northwest (cf. the red and black symbols between Figs. 3a and 3b and between Figs. 3c and 3d). The addition of moisture leads to a more intense and smaller-scale surface cyclone located more to the northeast (cf. Fig. 3a with Fig. 3c and Fig. 3b with Fig. 3d).

Figure 4 shows the northward and eastward displacement of the surface cyclones as a function of time by following both the low-level relative vorticity maximum and the anomalous SLP minimum. The total eastward displacement being largely dominated by the advection by the basic flow, its component has been subtracted from the total displacement to make the comparison between the simulations easier. A mean tendency of about 18 m s\(^{-1}\) has been subtracted corresponding to the basic flow velocity at 800 hPa in the area of cyclone travel. The cyclones in the two-perturbations simulations moves slower eastward than their one-perturbation counterparts (cf. the dashed and solid lines on Figs. 4a and 4b). After 60 h of simulation, the SLP minima in the one-perturbation runs are 150 and 200 km more to the west than those in the two-perturbations runs for the dry and moist simulations, respectively. Results for the vorticity maxima are roughly the same despite some discrepancies with SLP results. Note that these differences between the one and two-perturbations runs appear rather late, especially for the moist runs, which will be discussed later on. In contrast, the more rapid northward motion of the two-perturbations runs compared to the one-perturbation runs occurs much sooner at the early times, and the difference stays more or less constant with time after a while (Figs. 4c,d). The stronger difference appears for the moist runs. Differences between dry and moist runs are nonexistent in the beginning of the simulations, which is logical, as they have the same initial conditions, except for moisture, and it takes time for the moist processes to modify the dynamical features of the flow, which is creating this delay for achieving condensation. The faster eastward and northward motions in moist runs appear after a while, typically after 24–30 h, depending on the simulations. Differences in eastward displacement between the one-perturbation dry and moist cases are more or less the same as those between the two-perturbations dry and moist cases, on the order of 100 km for SLP minima and 200 km for vorticity maxima after 2 days.
In contrast, differences in northward displacement between dry and moist runs are almost 2 times larger in the two-perturbations cases than in the one-perturbation cases (cf. the red and black solid lines on one side and the red and black dashed lines on the other side in Figs. 4c, d). In the interval between $t = 30$ and $50$ h, the order of difference between dry and moist runs for SLP minima is about 50 and 100 km in the one- and two-perturbations runs, respectively. For vorticity maxima, it is about 100 and 200 km, respectively.

Our interpretation of the previously described cyclone tracks’ differences is mainly based on differences in shape and intensity of the upper-level anomalies that have their PV fields shown in Fig. 3. Whatever the simulations, an upper-level dipole composed of an upstream trough and a downstream ridge is centered around the surface cyclone. This dipole is created and/or reinforced through baroclinic interaction as the meridional velocities induced at upper levels by the surface cyclone PV advect the high background PV to the south on its upstream side and the low background PV to the north on its downstream side, generating a cyclone and an anticyclone, respectively (Hoskins et al. 1985). The other classical mechanistic argument related to baroclinic interaction is that the descending and ascending motions created by the surface cyclone in presence of the background vertical shear generate positive and negative relative vorticity anomalies in the upper levels (Holton 2004). Note that, in cases where there are no initial upper-level anomalies (Fig. 3), the upper-level anticyclonic PV is 3 times stronger in amplitude than the upper-level cyclonic PV. This can be explained by the so-called downstream development process (Chang et al. 2002). The background PV gradient being positive at upper levels, there is downstream dispersion of energy. In cases where the upper-level cyclone is present in the initial state, the two opposite-sign anomalies have more similar amplitudes after 48 h.
Because the upper dipole is southwest–northeast oriented, the velocities induced by the dipolar PV anomaly are mainly northwestward oriented. It is the more northwestward advection of the surface cyclone by the more intense dipole that explains the differences between the two-perturbations and one-perturbation cases shown in Figs. 3 and 4. The evolution of the dipole axis with time also explains why the differences in northward displacement are stronger at the early times and those in westward displacement occur much later. The cyclonic action of the surface cyclone onto the dipole acts to cyclonically turn the dipole axis, which renders it less and less zonally oriented with time. The advection by the dipole is thus more northward in the beginning and more westward at the end of the simulations.

Differences between the moist and dry runs can be partly explained by the intensity of the upper dipole. As the upper anticyclone is more intense in moist runs, it advects the surface cyclone faster to the north. However, the same argument cannot be applied to the zonal motion, as the surface cyclone translates faster eastward in moist runs than dry runs, despite having a stronger dipole that induces a more important westward advection. The stronger eastward translational speed in moist simulations is consistent with the theoretical and numerical studies on moist processes (Mak 1982; Davis et al. 1993). It is mainly attributed to the positive PV tendencies in low levels created by the released latent heat on the eastern side of the cyclone. This creates a cyclonic circulation tendency to the east of the cyclone that displaces it faster eastward. This eastward motion tendency is partly compensated by the effect of the upper anticyclone, which advects the surface cyclone more westward, at least at the later stage. Note that this effect is intrinsically nonlinear, as it needs the upper dipole to orient along the meridional direction more and more with time, which does not appear in linear simulations. In that sense, the present mechanism differs from that provided by de Vries et al. (2010), who pointed out,

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In the whole paper, the dipole orientation refers to the line joining cyclonic and anticyclonic centers of the dipole.
under the linear counterpropagating Rossby wave framework, the tendency of upper-level diabatically produced PV anomalies to slow down the eastward motion induced by the lower-level diabatically produced PV anomalies.

b. Interpretation using relative and potential vorticity budgets

The aim of the present section is to confirm the previous interpretations by making vorticity budgets.

1) RELATIVE VORTICITY BUDGET

To study the causes of the displacement of the upper-level relative vorticity and, thus, surface cyclone, the primitive-equations vorticity equation is decomposed as follows.

\[
\frac{\partial \zeta}{\partial t} + \nabla \cdot \mathbf{u} = -u' \frac{\partial \zeta}{\partial x} - v' \frac{\partial \zeta}{\partial y} - \omega \frac{\partial \zeta}{\partial z} - \frac{\partial}{\partial z} \left( \omega \frac{\partial \varphi}{\partial \zeta} \right) + \frac{\partial}{\partial \zeta} \left( \omega \frac{\partial \varphi}{\partial \varphi} \right) \mathbf{u} - (\zeta + f) \mathbf{v} 
\]  

(2)

The quantity \( A_1 = \frac{\partial \zeta}{\partial t} + \nabla \cdot \mathbf{u} \) corresponds to the vorticity tendency taking into account the advection by the basic flow; \( B_1 = -u' \frac{\partial \zeta}{\partial x} - v' \frac{\partial \zeta}{\partial y} \) is the nonlinear advection term; \( C_1 = -\omega \frac{\partial \zeta}{\partial z} - \frac{\partial}{\partial z} \left( \omega \frac{\partial \varphi}{\partial \zeta} \right) + \frac{\partial}{\partial \zeta} \left( \omega \frac{\partial \varphi}{\partial \varphi} \right) \mathbf{u} \) includes all the terms involving the vertical velocity (vertical advection + twisting term + vertical stretching); and \( D_1 \) is the dissipation term. The three tendency terms \( A_1, B_1, \) and \( C_1 \) are shown in Fig. 5 for the two moist simulations. Since the one-perturbation and two-perturbations runs do not have the same vorticity amplitudes, all the vorticity tendencies have been divided by their corresponding vorticity maxima to be comparable. For each simulation, the total vorticity tendency (divided by the vorticity maximum) exhibits large positive values over and slightly north of the vorticity maximum (Figs. 5a,b). There is thus both a reinforcement of the vorticity and a tendency to be displaced northward. The positive tendency is more important in the two-perturbations simulation than in the one-perturbation simulation, consistent with the fact that the vorticity maximum is more rapidly displaced northward in the former than in the latter. The nonlinear advection (Figs. 5c,d) shows a dipolar anomaly tendency that is roughly centered over the vorticity maximum and slightly east of it. For each simulation, the dipolar anomaly tendency is globally northward oriented but is stronger and contains a more important westward orientation in the two-perturbations simulation. Finally, the vertical term (Figs. 5e,f), which is dominated by the vertical stretching term, induces a positive tendency centered east of each vorticity maximum. Therefore, Fig. 5 shows that the reinforcement of the low-level vorticity is due to the vertical term and, in particular, the stretching term (not shown), while the northward displacement is explained by nonlinear advection. The stronger dipolar anomalies of the nonlinear advection term in the two-perturbations case compared to the one-perturbation case explains the faster northward displacement of the surface cyclone. It also explains why, at the end of the simulations, the two-perturbations case has a less-rapid eastward displacement than the one-perturbation case.

To determine how much the nonlinear advection tendencies are influenced by the upper-level disturbances, a PV inversion tool (see appendix) is used to differentiate low-level winds induced by the upper-level PV anomalies from the winds induced by the lower-level PV anomalies. Different nonlinear advection tendencies computed with different wind fields are shown in Fig. 6 for the moist simulation. The tendency computed with the wind field induced by all the PV anomalies over the whole atmospheric column is similar to the tendency computed with the real wind field, even though some slight discrepancies exist (cf. Fig. 5d with Fig. 6a). The southeast–northwest dipolar structure exists in both panels despite some slight variances. The tendency computed with the upper-level PV-induced wind field also shows a southeast–northwest dipolar structure, but with a stronger amplitude than the one induced by all the PV anomalies (cf. Figs. 6a and 6b). The tendency computed with the lower-level PV-induced wind field exhibits a north–south dipolar tendency, which has mostly the opposite effect to the upper-level PV anomalies. This is because of the fact that, at low levels, the positive PV maximum is located slightly to the east of the positive relative vorticity maximum, which tends to push the latter maximum southward. Note that the difference between the total PV and the upper-level PV-induced tendencies is similar, even though not equal, to the lower-level PV-induced tendency showing some degree of nonlinearity in the PV inversion algorithm (Figs. 6c,d). Since the upper-level PV effect gains the upper hand over the lower-level PV one, the net effect of nonlinear advection is a northwestward shift that can be mainly attributed to the wind field induced by the upper-level PV anomalies.

Figure 7 clearly shows that the stronger northwestward winds induced by the more intense upper-level PV anomalies in the two-perturbations runs compared to the one-perturbation runs (cf. Fig. 7a with Fig. 7b and Fig. 7c with Fig. 7d) explain the faster northward motion of the surface cyclone in the former runs, together with their less-rapid eastward motion. The stronger upper-level anticyclone in moist runs compared to dry runs also creates slightly more intense northwestward winds, but
differences are much less visible. A more representative computation of the upper-level PV-induced winds at the 850-hPa vorticity maximum is shown in Fig. 8. The westward component of the induced wind (Fig. 8a) is slightly stronger (less than 1 m s\(^{-1}\)) and the northward component (Fig. 8b) significantly stronger (about 2 m s\(^{-1}\)) in moist runs than in dry runs. A combination of two effects explains these differences. First, the
upper-level anticyclone is stronger in moist runs, leading to slightly stronger northwestward winds at low levels; second, the surface cyclone is closer to the upper-level anticyclone and, thus, to the strong winds the anticyclone induces at low levels in moist runs (cf. the position of the low-level cyclone relative to the upper-level anticyclone in Figs. 3b and 3d). Another interesting result of Fig. 8 concerns the time evolution of the orientation of the upper-level PV dipole and its effect on the induced winds at low levels. In the beginning of the simulations, the initial upper-level cyclone of the two-perturbations runs creates a net northward wind tendency in the low levels, which is absent in the one-perturbation runs, as it takes time for the latter runs to build up the upper anomalies. On the contrary, differences in upper-level PV-induced zonal winds between the two-perturbations and one-perturbation runs appear much later (after 36 h) at the time when the upper-level dipole axis becomes more zonally oriented and induces a more important westward wind (Fig. 8a). This explains why the difference in the eastward displacement is made progressively after 36 h, while, in the northward displacement, it is made in the first 24 h and then stabilizes.

To conclude, the nonlinear horizontal advection term is responsible for the northward displacement of the low-level vorticity maximum, while the vertical term, in particular the stretching term, explains the reinforcement of the vorticity and creates an eastward displacement tendency. This picture is valid for all simulations after some time (typically after 1 day) but does not necessarily reflect what is happening at the early stage of the simulation during which some spinup occurs (not shown).
2) SENSITIVITY TO VARIOUS POTENTIAL VORTICITY ANOMALIES

The potential vorticity budget developed here is also expressed to more clearly separate the different diabatic effects. The PV tendency by following the basic flow advection can be expressed as follows:

$$\frac{\partial Q}{\partial t} + \vec{u} \cdot \nabla Q = -u' \frac{\partial}{\partial x} Q - \nu \frac{\partial}{\partial y} Q - \omega \frac{\partial}{\partial p} Q + D,$$

where $Q$ denotes the PV. The total tendency $A_Q = \frac{\partial Q}{\partial t} + \vec{u} \cdot \nabla Q$, the nonlinear advection $B_Q = -u' \frac{\partial}{\partial x} Q - \nu \frac{\partial}{\partial y} Q$, and the vertical advection $C_Q = -\omega \frac{\partial}{\partial p} Q$ are computed using finite-difference schemes. The diabatic tendency $D$ is indirectly calculated from the other terms. These four potential vorticity tendencies are depicted on Fig. 9 for the two-perturbations moist simulation at $t = 48$ h. The total tendency (Fig. 9a) mainly presents positive values over the regions of high PV, with a peak occurring north of the PV maximum. This means that the tendency acts to reinforce the positive PV maximum and to shift it farther north. As in the relative vorticity budget, nonlinear horizontal advection presents dipolar anomaly tendencies that tend to displace the low-level PV maximum northwestward. Both the vertical advection (Fig. 9c) and the diabatic terms (Fig. 9d) have positive tendencies over the PV maximum region but reach their maximum amplitude to the east of the PV maximum. This means that both terms, but more importantly the diabatic term, act to intensify the positive PV maximum and to displace it eastward. To conclude on the PV budget, direct effects of moist processes represented by the diabatic term act to intensify the low-level PV maximum and to accelerate its eastward motion. There are also indirect effects, since the more intense upper-level anticyclone in the presence of moist processes reinforces the northward-tendency component of the nonlinear advection term. It confirms the idea that the cross-jet motion of the surface cyclone can
be interpreted as resulting from the nonlinear advection induced by the upper-level PV dipole which is reinforced in presence of moist processes.

c. Additional sensitivity experiments

To confirm the above findings and to more precisely identify the key moist processes acting on the cyclone motion, other sensitivity experiments were made. A first additional dry simulation was performed by initializing the model from the state of the moist run at 30 h, meaning that all physics related to the addition of moisture were disabled after 30 h. This simulation is called the moist-dry simulation. Both the eastward and northward motions of the surface cyclone are slowing down in the moist-dry simulation (cf. the blue curves with red ones in Figs. 10a,b). A 150-km difference in both the meridional and zonal direction is noticeable between the moist and moist-dry simulations after roughly 1 day. Once again, it shows that moist processes are responsible for an acceleration of both the eastward and northward motions of the surface cyclone.

A second additional simulation was performed with a simplified scheme using only transition phases between cloud water and water vapor, referred to as the cloud water simulation. In the cloud water simulation, the cyclone moves slower eastward and northward than in the moist simulation and faster than in the dry simulation. After 30 h of the simulation, when the differences in motion speed appear, the cyclone in the cloud water simulation is about 100 km behind the moist simulation cyclone and about 100 km ahead of the dry simulation cyclone in the east direction (green vs red and black lines on Fig. 10a). It is also about of the same order in the difference toward the north direction (green vs red and black lines on Fig. 10b). It can be noticed that the separation between the dry and cloud water simulations appears after 24 h of simulation, while the moist and cloud water simulations separate after 18 h of simulation. The position of the cyclone in the cloud water simulation at halfway between the dry and the moist simulation can be explained by the amplitude of the latent heat release, which is less important in the cloud water simulation than in the moist simulation. Less latent heat release leads to smaller positive PV tendencies under the heating zone and, thus, to a less-important eastward translational speed. It also means that the upper-level anticyclone of the cloud water simulation is weaker in amplitude favoring a less-rapid northward motion than in the moist run (not shown).

To confirm this difference in the intensity of latent heat release between the moist and cloud water simulations, a heating budget is shown on Fig. 11. Latent heat release in the moist simulation relies mostly on the water vapor deposition on snow, on the condensation and deposition on ice, and on the deposition on graupel (Fig. 11a). This heating occurs between the altitudes of 1000 and 7000 m. Note that the depositional growth of ice explains the peak of the blue curve at 4-km height, while condensation is responsible for the smaller peak of the same curve at 1-km height. In the cloud water simulation, the only processes that can occur are condensation and evaporation. As it happens here, it is the condensation that is at the origin of the latent heating, between 1000 and 6000 m above surface, and it corresponds to less than two-thirds of the latent heat release of the full moist simulation (red vs black lines on Fig. 11b).
Fig. 11b). This shows how the moist simulation produces more latent heat release than the cloud water simulation, thanks to the presence of the various cold processes mostly. Another important distinction between the cloud water and full moist simulations concerns the low levels. Below 1000 m, the heating rate $D\theta/Dt$ is negative in the moist simulation because there is sublimation of snow and also graupel melting and rain evaporation. It is zero in the cloud water simulation, because there is no precipitation that would allow a phase transition to water vapor. The vertical gradient of the heating rate is thus much more important in the full moist simulation, which leads to a more intense positive PV tendency at low levels and explains the faster eastward displacement of the surface cyclone.

4. Conclusions and discussion

The effects of upper-level disturbances and diabatic moist processes on the tracks of midlatitude idealized surface cyclones and, more precisely, their cross-jet motions have been systematically studied in the present study. The initial state of each idealized simulation included a finite-amplitude surface cyclone located to the south of a zonal baroclinic jet. In some runs, an upper-level cyclonic anomaly upstream of the surface cyclone was also initially added. Whatever the simulation, an upper-level dipolar anomaly centered over the surface cyclone and composed of an upstream cyclonic anomaly and a downstream anticyclonic anomaly was formed during the simulation. There was also a systematic crossing of the zonal jet by the surface cyclone after roughly 1–2 days. Vorticity budgets and the use of the PV inversion tool revealed that it is the nonlinear advection of the surface cyclone by the upper-level PV dipole that explains the cross-jet (or northward) motion of the surface cyclone, at least after waiting for some model spinup to occur. Because the dipole has a southwest–northeast tilted shape, it also tends to decelerate its eastward displacement. Sensitivities to the initial upper-level cyclone and moist processes led to the following results:

- The simulation with an initial upper-level cyclonic anomaly resulted in a stronger surface cyclone, a faster northward motion, and a slower eastward motion than the simulation without it, because it created a more
intense upper-level PV dipole. The time evolution of the dipole orientation also explains some nuances between the runs. Because the dipole axis was mainly meridional at the early stage and got a more southwest–northeast tilt at the later stage, the faster northward motion occurred in the former stage and the slower eastward motion in the latter.

- Comparison between moist and dry runs showed that the addition of moisture created a reinforcement of the surface cyclone and a faster northeastward motion. The faster northward motion is mainly explained by a more intense upper-level anticyclone due to the released latent heat. It is also the closer location of the surface cyclone to the strong upper-level PV-induced winds, which made it more sensitive to advection by the upper-level dipole in moist runs compared to dry runs. The faster eastward motion is due to the cyclonic circulation tendency at low levels created by the released latent heat, which overwhelms the westward tendency created by the more intense upper-level anticyclone. In the relative vorticity budget, the dominating eastward tendency is included in the terms involving the vertical velocity, such as the vertical stretching term, which is reinforced in presence of moisture, while in the PV budget it can be mainly attributed to the direct diabatic term. These differences of speed between the cyclones in the different cases make the crossing of the jet happen a few hours earlier with the upper-level perturbation or the moist processes.

Fig. 10. (a) Eastward and (b) northward displacement of the maximum relative vorticity anomaly averaged between 700 and 900 hPa as a function of time. The black and red solid curves correspond to the two-perturbations dry simulation and two-perturbations moist simulation, respectively. The green curve corresponds to the two-perturbations moist simulation with only cloud water and water vapor. The blue curve corresponds to the dry simulation initialized at \( t = 30 \) h with the initial conditions of the two-perturbations moist simulation.

Fig. 11. (a) Latent heat release budget made by averaging the different quantities over the whole domain at \( t = 48 \) for a moist simulation and (b) for cloud water simulation vs moist simulation.
Additional sensitivity experiments confirm these findings. In particular, the simulation in which only transition phases between cloud water and water vapor were permitted showed a slower northeastward motion than the moist simulation, in which all the water transition phases were included. Once again, the difference comes from the less-important latent heat release occurring in the former simulation than in the latter one.

This study is consistent with previous ones concerning the intensification of the surface cyclone and the effects of moist processes on the accentuation of the eastward motion (e.g., Davis et al. 1993). The more original aspect of our study relies on a detailed description of moist processes acting on the latitudinal displacements of surface cyclones. As in the idealized quasigeostrophic processes on the accentuation of the eastward motion (e.g., Davis et al. 1993). The more original aspect of our study relies on a detailed description of moist processes acting on the latitudinal displacements of surface cyclones. As in the idealized quasigeostrophic simulations of Gilet et al. (2009) or Oruba et al. (2013), and in the real storm simulations of Rivière et al. (2012), it is the upper-level PV dipole that is shown to play a key role in the cross-jet motion of the surface cyclone. In the present study, moist processes are shown to intensify the downstream ridge of the dipole and to accelerate the cross-jet motion. In that sense, the present study confirms the basic theory described in simple dry quasigeostrophic models and shows how much moist processes act to accentuate dry dynamics, even though they do not form a new mechanism, per se. In the future, an interesting point to study would be to analyze real cases of jet-crossing cyclones and to quantify the role of the different moist processes in their motion speed.

APPENDIX

Potential Vorticity Inversion

Before presenting the inversion algorithm itself, a scaling analysis is made as in Guérin et al. (2006).

a. Potential vorticity scaling

Following Hoskins and Bretherton (1972), we define a pseudoheight $z^*$ and a pseudodensity $\rho^*$:

$$z^* = \frac{C_p}{g T(p_0)} \left[ 1 - \left( \frac{p}{p_0} \right)^{R C_p} \right]$$ and

$$\rho^* = \rho_0 \left( \frac{p}{p_0} \right)^{1 - R C_p},$$

where $\theta_0 = T(p_0)$ is the temperature at $p_0 = 1000$ hPa in a motionless, isentropic atmosphere, $R$ is the gas constant, $C_p$ is the specific heat at constant pressure for dry air, $\rho_0$ is the density at level $p_0$, and $g$ is the acceleration because of gravity. The potential vorticity $Q$ is written using the pseudoheight $z^*$:

$$\rho^* Q = (f + \xi) \frac{\partial \theta}{\partial z^*} + \frac{\partial u}{\partial z^*} \frac{\partial \theta}{\partial y} - \frac{\partial v}{\partial z^*} \frac{\partial \theta}{\partial x}. \quad (A1)$$

The wind and temperature are assumed in thermal wind balance:

$$f_0 \frac{\partial u}{\partial z^*} = \frac{g}{\theta_0} \frac{\partial \theta}{\partial x}$$

and

$$f_0 \frac{\partial v}{\partial z^*} = -\frac{g}{\theta_0} \frac{\partial \theta}{\partial y}.$$

At this point, a nonconventional temperature splitting is proposed since it differs from the one leading to the quasigeostrophic governing equations:

$$\theta(x, y, z^*) = \overline{\theta}(z^*) + \theta_1(x, \epsilon y, z^*) + \theta_2(x, \epsilon y, z^*),$$

where $\epsilon$ is small with respect to one. The decomposition is motivated by the fact that the local environmental static stability in which a PV anomaly is embedded can be significantly different from the horizontal mean $\overline{\theta}(z^*)$. We therefore introduce a second reference state ($\theta_1$) slowly varying with $x$ and $y$. Since $\epsilon$ is small with respect to one, the contribution of the horizontal gradients of $\theta_1$ in the thermal wind balance are negligible. Besides, the vertical gradient of $\theta_1$ can be strong and comparable to that of $\overline{\theta}(z^*)$. Thus, a reference static stability will be used for the scaling of both $\overline{\theta}$ and $\theta_1$:

$$\rho^* Q = (f + \xi) \left( \frac{\partial \overline{\theta}}{\partial z^*} + \frac{\partial \theta_1}{\partial z^*} + \frac{\partial \theta_2}{\partial z^*} \right)$$

$$- \frac{f \theta_0}{g} \left[ \left( \frac{\partial u}{\partial z^*} \right)^2 + \left( \frac{\partial v}{\partial z^*} \right)^2 \right]. \quad (A2)$$

Each variable $\mathcal{M}$ can be written as $\mathcal{M} = [\mathcal{M}] \mathcal{M}'$ where $[\mathcal{M}]$ is a scaling of the corresponding variable, and $\mathcal{M}'$ is the dimensionless variable. The scaling for the subset of independent variables is as follows:

$$[U] = U, \quad [\delta x] = [\delta y] = L, \quad [\delta z^*] = H,$$

$$N^2 = \frac{g \theta_0}{\theta_0} \left( \frac{\partial \overline{\theta}}{\partial z^*} \right)^2,$$

where $N$ is the Brunt–Väisälä frequency of the reference state. We then have

$$[\xi] = U/L, \quad [\partial U/\partial z^*] = U/H.$$

The two temperature perturbations are not scaled along the same lines. The first one is comparable to the reference static stability. Therefore, $[\partial \theta_1/\partial z^*] = (\theta_0/g) N^2$. The
second one is scaled using the thermal wind balance and leads to \( \frac{\partial \theta}{\partial z} = \theta / g H \).

A Rossby number \( Ro = U/L \) and Froude number \( F = U/NH \) are introduced. The dimensionless form of potential vorticity becomes the following:

\[
\frac{\rho^*}{f_0 N^2} \frac{\partial}{\partial z} \left[ Q \right] = \frac{\partial \theta}{\partial z} + \frac{\partial \theta^f}{\partial z} + Ro^{-1} F^2 \frac{\partial^2 \theta^f}{\partial z^2}
\]

\[
+ Ro \frac{\partial \theta}{\partial z} + Ro \frac{\partial \theta^f}{\partial z} + F^2 \frac{\partial^2 \theta^f}{\partial z^2}
\]

\[
- F^2 \left( \frac{\partial u}{\partial z} \right)^2 - \left( \frac{\partial v}{\partial z} \right)^2 \quad \text{(A3)}
\]

b. Potential vorticity truncation

We then choose \( Ro = F = 0.1 \). Therefore, if we keep the term \( \xi \frac{\partial \theta}{\partial z} \) in the PV formulation [Eq. (A1)], the scaling shown in Eq. (A3) shows that we are retaining all the first-order terms but also some second-order ones. The present truncation may be thus considered as peculiar, since the terms in \( F^2 \) are not treated in a consistent way. Note also that some terms, cancelled here and related to ageostrophy, are at least \( F^2 Ro \), as shown in Guérin et al. (2006).

With dimension, the potential vorticity that is inverted is then

\[
\rho^* Q = (f + \xi) \frac{\partial \theta}{\partial z} \quad \text{(A4)}
\]

c. Inversion

Here, inverting the potential vorticity means finding the wind and temperature fields in thermal wind balance that obey the PV equation [Eq. (A4)]. The present formulation is nonlinear and cannot be inverted at once like within the quasigeostrophic (QG) framework. Therefore it is reasonable to use a decomposition between the nonlinear and linear terms, as suggested above and similarly by Arboagast et al. (2008):

\[
\rho^* Q - \xi \left( \frac{\partial \theta}{\partial z} + \frac{\partial \theta^f}{\partial z} \right) = \left( f \frac{\partial \theta}{\partial z} + \xi \frac{\partial \theta^f}{\partial z} \right) \quad \text{(A5)}
\]

Using geostrophic balance, the left-hand side is a nonlinear function of the geopotential \( \phi \), whereas the right-hand side becomes the sum of the vertical derivative and the horizontal Laplacian of \( \phi \). Thus, it can then be written as a 3D Laplacian:

\[
\rho^* Q - F(\phi) = \Delta_3 \phi \quad \text{(A6)}
\]

This equation is solved iteratively following a fixed-point strategy for \( F \). The left-hand side is fixed during the Laplacian inversion. A provisional solution is found and is then introduced in the left-hand side. The Laplacian is inverted again with the new source term, and so on. Formally, the iteration is:

\[
\rho^* Q - F(\phi_n) = \Delta_3 \phi_{n+1}
\]

where \( n \) is the iteration number.

The solution of this problem is unique when a boundary condition is specified. Here, a Neumann condition for \( \phi \) is prescribed. We therefore specify \( \phi \) at the upper and lower boundaries of the inversion domain.

Since the inversion problem is nonlinear, the attribution of a geopotential anomaly to a PV anomaly is not unique. We proceed along the following lines: we consider a known atmosphere state (a model output) defined by its PV, \( Q_n \), and its boundary, \( \theta^f_n \). We define anomalies \( Q_A \) and \( \theta^f_A \), and compute the new geopotential \( \phi_n \) associated with the model PV, \( Q_n = Q_M - Q_A \) and the new boundary condition \( \theta^f_n = \theta^f_M - \theta^f_A \).

The new geopotential \( \phi_n \) then obeys the following equation:

\[
\rho^* Q - F(\phi_n) = \Delta \phi_n \quad \text{(A7)}
\]

The difference \( \phi_M - \phi_N \) can be considered as the component of the flow associated to the PV anomaly \( Q_A \).

REFERENCES


