Physical Mechanisms of Tropopause Polar Vortex Intensity Change

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ABSTRACT

Characterized by radii as large as 800 km and lifetimes up to months, cyclonic tropopause polar vortices (TPVs) are coherent circulation features over the Arctic that are important precursors for surface cyclogenesis in high and middle latitudes. TPVs have been shown to be maintained by radiative processes over the Arctic owing to limited amounts of latent heating. This study explores the hypothesis that a downward extension of dry stratospheric air associated with TPVs results in an increase in longwave radiative cooling that intensifies the vortex.

Idealized numerical modeling experiments are performed to isolate physical interactions, beginning with radiative forcing in a dry atmosphere and culminating with multiple physical interactions between radiation and clouds that more accurately represent the observed environment of TPVs. Results show that longwave radiative cooling associated with a rapid decrease in water vapor concentration near the tropopause is primarily responsible for observed TPV intensification. These enhanced water vapor gradients result from a lower tropopause within the vortex that places dry stratospheric air above relatively moist tropospheric air. Cloud-top radiative cooling enhances this effect and also promotes the maintenance of clouds by destabilizing the region near cloud top. Shortwave radiation and latent heating offset the longwave intensification mechanism. Heating from shortwave radiation reduces the cloud water mixing ratio by preferentially warming levels above cloud tops.

1. Introduction

Radiation is a key component for the maintenance and intensification of cyclonic tropopause polar vortices (TPVs) over the Arctic (Cavallo and Hakim 2010, 2012). TPVs are tropopause-based coherent vortices that are frequently observed poleward of the midlatitude jet stream (e.g., Hakim and Canavan 2005), and play an important role in extratropical cyclogenesis when captured by the jet stream (e.g., Hakim et al. 1995; Bosart et al. 1996; Hakim et al. 1996). However, TPVs can exist on relatively long time scales before transitioning from higher- to lower-latitude regions, occasionally having lifetimes on the order of months. Cavallo and Hakim (2010, 2012) show that there is a net tendency for radiative processes to intensify TPVs, with qualitative evidence suggesting that the observed radiative vortex intensification is a response to changes in relative humidity associated with a characteristic downward intrusion of dry stratospheric air in the vortex core. Here, we examine the dynamics of idealized simulations of TPVs to better understand the physical mechanisms leading to their observed intensification.

Motivation to understand the maintenance and longevity of TPVs originates from a predictability standpoint since TPVs are characterized by an upper-level potential vorticity (PV) anomaly, an important feature necessary for type B cyclogenesis (e.g., Petterssen and Smeye 1971). Surface cyclogenesis is more likely to occur when upper-level PV anomalies move over regions of enhanced low-level baroclinicity; therefore, upper-level PV anomalies can be viewed as precursors to surface cyclogenesis (e.g., Hoskins et al. 1985). This can occur remotely in lower latitudes when TPVs are advected equatorward by the jet stream, or locally over the Arctic where recently a TPV led to the development and rapid intensification of an unusually strong surface cyclone over the Arctic Ocean, subsequently impacting the sea ice distribution (Simmonds and Rudeva 2012). Upper-level PV anomalies are frequently observed to originate over the Arctic, which equates to TPVs when they spend a majority of their lifetimes over the Arctic (Hakim and Canavan 2005). The vortical rather than wavelike property of TPVs (e.g., Hakim 2000), together

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with its conservation properties, makes PV a particularly useful diagnostic for quantifying TPV intensity change. For example, Cavallo and Hakim (2009) use a mesoscale numerical model to quantify the intensity change of a TPV over the Canadian Arctic, and showed that radiation was responsible for the intensification of that particular TPV. Cavallo and Hakim (2010) composited 568 numerically simulated TPV vortices over a fixed region of the Canadian Arctic and found a net tendency for radiative processes to strengthen the vortex. Latent heating has a smaller weakening effect, while all other diabatic processes were negligible.

Here we build upon this earlier work to isolate the dynamical consequences of various thermodynamic processes on affecting TPV intensity. Specifically, we compare the physical processes associated with infrared and solar radiation, and from latent heating. We proceed with a review of the methods used to diagnose vortex intensity change and a description of the numerical model and experimental setup in section 2. Results of the numerical experiments will be presented in section 3, with an examination of the dynamical and physical mechanisms leading to TPV intensification. A summary and conclusions of these results will be discussed in section 4.

2. Methods

a. Vortex intensity change

We define vortex intensity using Ertel potential vorticity (EPV, \( \Pi \)) (e.g., Pedlosky 1992):

\[
\frac{D \Pi}{Dt} = \frac{\omega_a}{\rho} \cdot \nabla D \theta + \frac{\nabla \theta}{\rho} \cdot \left( \nabla \times \frac{\mathbf{F}}{\rho} \right),
\]

where

\[
\Pi = \frac{\omega_a}{\rho} \cdot \nabla \theta.
\]

Above,

\[
\frac{D}{Dt} = \frac{\partial}{\partial t} + \mathbf{u} \cdot \nabla + \frac{\partial}{\partial y} + \frac{\partial}{\partial z}
\]

is the time rate of change following the fluid; the three-dimensional absolute vorticity vector is \( \omega_a = \omega + 2 \Omega = \nabla \times \mathbf{U} + 2 \Omega \), where \( \Omega \) is Earth’s rotational vector; \( \rho \) is the fluid density; \( \nabla = (\partial/\partial x, \partial/\partial y, \partial/\partial z) \) is the gradient operator; and \( \mathbf{U} = (u, v, w) \) is the three-dimensional wind vector; \( \theta = T(p_o/p)^{Rc_p} \) is the potential temperature, where \( T \) and \( p \) are temperature and pressure, respectively, \( p_o = 10^5 \text{ Pa} \) is a standard constant, \( R = 287 \text{ J K}^{-1} \text{ kg}^{-1} \) is the dry air gas constant, and \( c_p = 1004 \text{ J K}^{-1} \text{ kg}^{-1} \) is the specific heat capacity of dry air at constant pressure; and \( \mathbf{F} \) is the frictional force vector.

As shown by Cavallo and Hakim (2009, 2010), in the case of TPVs, (1) can be reasonably well approximated by neglecting \( \mathbf{F} \) since it is comparatively small near the tropopause:

\[
\frac{D \Pi}{Dt} \approx \frac{\omega_a}{\rho} \cdot \nabla \theta_{\text{longwave}} + \frac{\nabla \theta_{\text{shortwave}}}{\rho} + \frac{\nabla \theta_{\text{latent heating}}}{\rho} + \frac{\nabla \theta_{\text{other}}}{\rho},
\]

where \( \theta_{\text{longwave}} \) is the longwave radiative potential temperature tendency; \( \theta_{\text{shortwave}} \) is the shortwave radiative potential temperature tendency; \( \theta_{\text{latent heating}} \) is the potential temperature tendency due to latent heating; and \( \theta_{\text{other}} \) is the potential temperature tendency from all other diabatic tendencies including those from convection, planetary boundary layer effects, and tendencies resulting from explicit mixing processes. In the following section, we describe the experimental design and numerical model, where each of the above tendencies can be methodically isolated in an idealized setting in order to separate the effects of these processes from one other.

b. Experimental setup and design

Five numerical simulations are designed to systematically examine the intensification mechanisms of TPVs defined by the rhs of (3) (Table 1). In the first two experiments, denoted as E1 and E2, only the effects of longwave radiation are considered as a source for vortex intensity change. E2 is identical to experiment E1 except there is no water vapor in E1. Shortwave radiation is added in experiment E3. In experiment E4, the interactive effects between longwave radiation and latent

<table>
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<tr>
<th>Expt</th>
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<tr>
<td>E1</td>
<td>LW, without H2O</td>
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<tr>
<td>E2</td>
<td>LW, with H2O</td>
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<td>E4</td>
<td>LW, MP</td>
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<td>E5</td>
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TABLE 1. Summary of numerical simulation parameters longwave (LW) and shortwave (SW) radiation and microphysics (MP).
heating are examined. The combined effects of short-wave radiation, longwave radiation, and latent heating are considered in experiment E5.

c. Arctic vortex initialization

The initial vortex is specified using the Three-Dimensional Vortex Perturbation Analysis and Simulation (3DVPAS) (see Nolan and Montgomery 2002; Nolan and Grasso 2003; Hodyss and Nolan 2007; Nolan et al. 2007), with details extracted to appendix A. In the experiments inclusive of water vapor (E2–E5), water vapor is initialized such that relative humidity is 100% everywhere in the troposphere, with a gradual reduction to a fixed background stratospheric value above the tropopause. This is done to create an initial state in which there is a dry air anomaly owing to a lowered tropopause while avoiding a transition state before cloud formation; see appendix A for further details. To insure that the initial vortex is dynamically stable, the system is expressed in the form of a linear dynamical system:

\[ \frac{d}{dt} \mathbf{x} = T \mathbf{x}, \]  

(4)

where \( T \) is a constant matrix and \( \mathbf{x} \) is a state vector containing \( u, v, w, \) and \( \theta \) on each grid point. The eigenvalues of \( T \) are then solved to obtain the growth rates, which are less than zero if the system is dynamically stable. Figure 1 shows the growth rates for the first eight azimuthal modes of the system described above, reflecting a dynamically stable vortex. Given a dynamically stable axisymmetric initial vortex, what remains is to interpolate the vortex into the model coordinates of the numerical model described subsequently.

d. Numerical model

We use the Advanced Research Weather (ARW) version of the Weather Research and Forecasting Model (WRF), version 3.1.2 (Skamarock et al. 2008). Time integration is achieved using a third-order accurate Runge–Kutta method, and horizontal advection by a fifth-order accurate scheme. A horizontal grid spacing of \( \Delta x = \Delta y = 24 \) km is used with \( 200 \times 200 \) horizontal grid points, 61 vertical levels, and a model time step of \( \Delta t = 120 \) s. The following physics schemes are used: Rapid Radiative Transfer Model longwave radiation (Mlawer et al. 1997), the National Aeronautics and Space Administration Goddard shortwave radiation (Chou and Suarez 1994), and WRF single-moment 5-class microphysics (WSM5; Hong et al. 2004). No surface, surface layer, boundary layer, or cumulus physics schemes are used, and the surface is set to sea level. Simulations are performed on an \( f \) plane centered at 75°N, and the surface is set to be ice with an initial surface skin temperature equal to the air temperature at a height of 0 m. The surface skin temperature is used as the radiative lower boundary condition. Horizontal boundaries are periodic, and there is no upper-level diffusive damping, vertical velocity damping, or hyperdiffusion. To eliminate solar differential heating effects, the effective latitude for solar radiative effects is fixed to 90°N, and the time of day is fixed at solar noon at all grid points. All simulations are initialized 1 July and are run 120 h perpetually on this date unless otherwise stated. After the initial vortex is interpolated from the radius–height (\( r-z \)) grid of 3DVPAS to the WRF grid, an additional balancing procedure is employed as described in appendix B.

3. Results

The initial state of the TPV used for all experiments is summarized in Fig. 2. The downward extension of the dynamic tropopause is evident, from background values of around 375 to 600 hPa in the vortex core (Fig. 2a). EPV anomalies reach a maximum of \( \sim 5 \) PVU (1 PVU = \( 1 \times 10^{-6} \) K kg\(^{-1}\) s\(^{-1}\) m\(^2\)) in the lower-stratospheric portion of the vortex where the greatest anomaly in the vertical stratification of \( \theta \) and positive absolute vertical vorticity correlation exists. Potential temperature anomalies are negative (positive) below (above) the tropopause.
in the vortex core, while geopotential height anomalies are negative and centered at the background tropopause level in the vortex core (Figs. 2b,c). The tangential winds reflect the cyclonic circulation associated with the negative (positive) geopotential (EPV) anomalies in the vortex, with a maximum of 13 m s\(^{-1}\) at a radius of \(~500\) km from the vortex core (Fig. 2d). Negative relative humidity anomalies of \(~60\)% are present in the lower-stratospheric
portion of the vortex core (Fig. 2e). Background ozone mixing ratios are set using the subarctic winter (SAW) standard profile, from the Intercomparison of Radiation Codes Used in Climate Models (ICRCCM) (Ellingson et al. 1991). In this single-column profile, the tropopause is located at the pressure level where temperature no longer decreases with height, as defined by the World Meteorological Organization (1992). The pressure at this tropopause level is then used to define tropopause-relative ozone mixing ratios for each grid point in the domain. Figure 2f shows that ozone mixing ratios are locally greater in the vortex core with respect to the background environment, with maximum positive ozone anomalies of about 1 ppmv near 150 hPa. We proceed to the experiments described in Table 1, beginning with longwave radiative forcing without the effects of water vapor.

a. Experiment E1: Longwave radiative forcing exclusive of water vapor

For this experiment, changes in vortex strength are determined from term A in (3). Water vapor mixing ratios are set to zero everywhere in the domain, so the longwave radiative effects are due to temperature, carbon dioxide, and ozone. Referring to Fig. 2b, considering temperature only, we expect that positive (negative) potential temperature anomalies will roughly result in relatively greater (less) longwave cooling in the vortex core with respect to the background environment.

![Fig. 3. Time series plots of vortex amplitude for expts E1–E5 as labeled in the right margin. Amplitudes are computed on the 2-PVU surface.](image)

![Fig. 4. (a) Time series of vortex amplitude, and time–height sections of (b) temperature, (c) radiative heating rates, and (d) total EPV tendencies in expt E1 days 0–240. The solid black contour denotes the 2-PVU surface at the grid point closest to the vortex core. The color interval is 2 K in (b), 0.02 K day$^{-1}$ in (c), and 0.001 PVU day$^{-1}$ in (d). The zero contour in (b)–(d) is denoted by a solid gray contour. All values in (b)–(d) are averaged within a 500-km radius of the vortex core.](image)
In other words, vertical slopes of the isentropes should flatten over time, implying that the vortex will weaken. Figure 3 shows the change of vortex amplitude over the 5-day simulation period, where amplitude is the absolute difference between $\theta$ at the vortex core and the maximum $\theta$ on the 2-PVU surface. Contrary to expectation, vortex amplitude steadily increases at a nearly constant rate from 23.75 to 25 K. Extending the simulation length to 240 days reveals that the vortex amplitude increases until day $\sim$150 when a maximum amplitude of nearly 34 K is obtained (Fig. 4a).

This surprising result derives from the fact that the initial vortex has finite amplitude. To understand this aspect, we expand the longwave radiative heating term in (3) into two contributions: one from the surrounding environment, $\dot{\theta}_b$, and one within the vortex, $\dot{\theta}_v$, so that

$$\frac{DI_{\text{longwave}}}{Dt} \sim \frac{\omega_p}{\rho} \cdot \nabla(\dot{\theta}_b) + \frac{\omega_p}{\rho} \cdot \nabla(\dot{\theta}_v). \quad (5)$$

In the term involving $\dot{\theta}_v$, $\partial \dot{\theta}_v/\partial z$ is negative inside the vortex since the warm anomaly radiatively cools faster than the cold anomaly, and therefore this term weakens the vortex (not shown). However, in the term involving $\dot{\theta}_b$, $\partial \dot{\theta}_b/\partial z$ is positive both inside and outside of the vortex. At earlier times, when the vortex is strengthening, the warmest air and strongest longwave cooling is located at the top of a temperature inversion near 850 hPa (Figs. 4b,c). At a given level in the troposphere, temperature increases away from the vortex, so that $\partial \dot{\theta}_b/\partial z$ also increases away from the vortex. In the environment, EPV is generated at the rate $\sim \rho^{-1} \partial \dot{\theta}_b/\partial z$, whereas in the vortex EPV is generated at the rate

$$\frac{\zeta + f \partial \dot{\theta}_b}{\rho} \frac{\partial}{\partial z}$$

in which $\zeta$ is the vertical component of relative vorticity and $f = 2\Omega \sin(\text{lat})$ is the Coriolis parameter. Thus, if the relative vorticity is sufficiently large, as is the case here, then EPV is created preferentially inside the vortex. After the vertical gradient in radiative cooling reverses, the stronger vorticity inside the vortex instead acts to preferentially destroy EPV inside the vortex and weaken it (Figs. 4c,d). The experiment here shows that the time scale for the reversal of the vertical radiative cooling gradient is $\sim$150 days.

b. Experiment E2: Longwave radiation and the influence of water vapor

Experiment E2 is identical to E1 in all regards except for the presence of water vapor. Figure 3 shows that the vortex strengthens an additional 2.8 K in the 5-day simulation as compared to E1, or by 16.8%. Figure 5 shows cross-vortex sections of the average heating rates and EPV tendencies over the 5-day experimental period. Heating rates and EPV tendencies are quite different than in experiment E1 and, in particular, show enhanced radiative cooling near the tropopause with locally reduced longwave cooling in the vortex core above the tropopause (Fig. 5a). Since water vapor is a strong absorber of longwave radiation and generally decreases upward, it typically contributes to longwave cooling. This effect can be seen when comparing longwave cooling rates to the vertical gradients of water vapor in the troposphere (Fig. 6). Vortex intensification occurs when longwave cooling below the tropopause is locally stronger in the vortex core than in its surroundings, and is shown in Fig. 6b. At the initial time, both the minimum in the vertical gradient of water vapor and maximum longwave cooling rate anomalies are located $\sim$100–150 hPa below the tropopause. After 5 days, while the magnitudes of
both vortex anomalies remain nearly the same, vertical mixing of water vapor leads to an increase in the vertical extent of relatively high negative vertical gradient in water vapor. In response, the strongest longwave cooling rates also expand in vertical extent and shift upward to the tropopause. Reduced longwave cooling in response to the negative temperature anomaly near 600–850 hPa is offset by the large negative vertical gradient in water vapor near 700 hPa, resulting in a longwave cooling minimum around 700 hPa. This is particularly true in the vortex core, as the lowered tropopause increases the vertical gradient in water vapor in the troposphere since the mixing ratio decreases to stratospheric values over a thin layer.

Above the tropopause, in the vortex core, water vapor concentrations are lower than for surrounding locations at the same pressure level. Thus, the vertical gradient in water vapor is anomalously high, resulting in anomalous
longwave cooling. The combination of anomalously strong longwave cooling just below the tropopause and anomalously weak longwave cooling just above the tropopause in the vortex core results in a positive Ertel potential vorticity tendency near the tropopause in the vortex core (Fig. 5b).

The EPV tendency pattern in Fig. 5b here is quantitatively similar to the composite EPV tendencies of tropopause polar vortices over the Canadian Arctic documented by Cavallo and Hakim (2010). Thus, we conclude that the longwave radiative effects associated with vortex water vapor anomalies are important for TPV intensification.

c. Experiment E3: Inclusion of shortwave radiation

For this experiment, changes in vortex amplitude are determined from terms A and B in (3). Results show that the amplitude for experiment E3 evolves very similarly to E2 except that the overall vortex intensification is slightly less, ending with an amplitude of 27.3 K after 5 days (Fig. 3). The net heating rates and EPV tendencies are also similar to experiment E2, except that the net cooling rates are slightly weaker for E3, leading to a slightly smaller magnitude of EPV creation in the vortex core (Figs. 7a,b). Shortwave radiative heating rates are a maximum (minimum) 100 hPa below (above) the tropopause (Fig. 7c). Above ∼300 hPa, shortwave radiative heating rates increase with height. In the troposphere, shortwave radiative heating rates closely resemble the vertical gradients in water vapor, similar to what is seen for longwave radiation as discussed in section 3b for experiment E2. However, shortwave radiation is not as strongly absorbed by water vapor as longwave radiation, and hence the magnitudes of the shortwave heating near the tropopause are considerably smaller than the magnitudes of longwave cooling.

FIG. 7. Time-mean cross-vortex sections of the (a) total heating rate ($\dot{\theta}$), (b) total EPV Lagrangian tendency \((D\theta/Dt)\), (c) shortwave radiative heating rate ($\dot{\theta}_{\text{shortwave}}$), and (d) EPV Lagrangian tendency due to shortwave radiation \((D\theta/Dt)_{\text{shortwave}}\) for expt E3. The thick black contour denotes the 2-PVU surface. The zero contour is denoted by the thin solid contour. The color interval is 0.1 K day\(^{-1}\) in (a) and (c) and 0.02 PVU day\(^{-1}\) in (b) and (d).
the tropopause contributes to a negative EPV tendency (Fig. 7d).

As just demonstrated, near the tropopause shortwave heating from water vapor slows intensification. However, radial gradients in ozone may also contribute to such warming, so to test the sensitivity to ozone an additional experiment is performed that is configured the same as E3 but without ozone. The differences in vortex amplitude are small (<0.2 K) over the entire 5-day simulation (not shown). Radiative heating rate differences become more pronounced with height in comparison to E3; however, there remains little variation in vertical heating rate gradients near the tropopause.

Together, experiments E1–E3 show the radiative impact on TPV intensification is primarily due to enhanced vertical water vapor gradients near the tropopause. The greatest response to shortwave radiation is also from water vapor; however, this effect only slightly offsets the longwave radiative intensification.

d. Experiment E4: Longwave radiation and condensation of water vapor

For this experiment, changes in vortex amplitude are determined from terms A and C in (3). Results show that vortex intensification is greater in E4 than in any other experiment; vortex amplitude increases to 31.75 K after 5 days (Fig. 3). Compared to E2, time-mean (averaged over the entire 120-h simulation) net cooling rates are stronger at most levels, especially in the lower- and midtropospheric levels, resulting in more EPV creation in the upper troposphere (Figs. 8a,b). Longwave radiative heating rates are nearly the same as net heating rates, but differ considerably from those in E2 (cf. Figs. 5a, 8c). In particular, the greatest longwave cooling is shifted from near the tropopause level in E2 to around 700–800 hPa in E4 (Fig. 8c). Similar to E2, a longwave cooling minimum exists ~100 hPa above the tropopause in the vortex core, resulting in a similar decrease of longwave cooling with
height and stronger EPV creation owing to longwave radiation in the vortex core than in the background environment (Fig. 8d). Average latent heating rates of around 0.5 K day$^{-1}$ are confined mainly to the lower troposphere from about 600 to 900 hPa, with negligible EPV tendencies directly due to latent heating in comparison to longwave radiation (not shown).

More EPV is created in the vortex core in E4 than in E2, despite that latent heating destroys PV (Fig. 9a). Additional EPV creation from longwave radiation is a result of enhanced longwave radiative cooling in areas where strong vertical cloud$^3$ gradients are present (Figs. 9b,c). This increases the vertical longwave radiative heating gradient in those areas relative to E2, creating more upper-tropospheric EPV than would otherwise have been created in the absence of clouds. The impact of clouds on radiative forcing is illustrated by performing another experiment identical to E4, except cloud radiative forcing is removed. For this case, the EPV tendency (Fig. 9d) reveals that the additional EPV creation in the upper troposphere is indeed due to enhanced longwave radiative cooling near cloud top associated with cloud-radiative feedbacks.

The environmental conditions in this experiment most closely resemble the true environmental conditions surrounding TPVs in the Arctic during winter since shortwave radiation is seasonally limited. Furthermore, clouds are more likely to be associated with TPVs since they are associated with reduced

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$^3$Cloud in this manuscript refers to the sum of cloud ice and cloud liquid water mixing ratios.
static stability and higher relative humidities in the lower troposphere. While experiment E2 showed that the intensification observed by Cavallo and Hakim (2010) over a large sample of TPVs can largely be reproduced by accounting for anomalous vertical water vapor gradients in the vortex core, the combined effects of water vapor and cloud-radiative effects seen in E4 supports an enhancement in TPV intensification from clouds, as seen by Cavallo and Hakim (2009) during a particular event.

**e. Experiment E5: Longwave and shortwave radiation, inclusive of latent heating**

For this experiment, changes in vortex amplitude are determined from terms A, B, and C in (3). Vortex amplitude follows a similar pattern as in experiment E4 for approximately the first two days except the peak amplitude within this 2-day period occurs earlier, with a slightly lower amplitude of 26.1 K rather than 27.8 K (Fig. 3). After 2 days, the vortex amplitude more closely resembles experiments E2 and E3, and at the end of the 5-day simulation reaches 28.2 K.

Figure 10 shows the time-mean heating and EPV tendencies for experiment E5. In comparison to E4, there is less radiative cooling, especially above the tropopause (cf. Fig. 8a and 10a). Differences in the net time-mean EPV tendencies are similar to E4 at pressures greater than ~300 hPa, with slightly more (less) EPV creation just below (above) the vortex core tropopause (cf. Figs. 8b and 10b). Time-mean longwave radiative heating and the corresponding EPV tendencies owing to longwave radiation are also very similar to E4, with slightly more EPV creation below the vortex core tropopause (not shown). Shortwave radiative heating rates are similar to those found in experiment E3 except that the maximum heating rates are lower in magnitude and are shifted to ~700 from ~500 hPa (cf. Figs. 7c and 10c). The 700-hPa level is near the mean cloud-top level in this simulation. Just above the clouds, vertical gradients in water vapor
are smaller than they otherwise would be without cloud formation below because water vapor in the levels below has been converted into cloud liquid water or ice. This reduces longwave cooling rates but increases shortwave heating rates owing to the radiative effects from changes in water vapor. Cloud formation and latent heating occur well below the tropopause and at magnitudes lower than the corresponding longwave and shortwave radiative heating rates. Although clouds form both inside and away from the vortex, latent heating rates are greater away from the vortex owing to warmer temperatures. The resulting EPV tendencies are small in comparison to the radiative EPV tendencies (not shown).

Figure 11 shows the time evolution of differences in the EPV and radiative EPV tendencies between experiments E4 and E5 in the vortex core. Overall, EPV generation is less near the tropopause in the vortex core, as expected, since intensification is slower in E5 than in E4 (Fig. 11a). Mean cloud mixing ratios are lower in E5 with shortwave radiation, except briefly around 80–90 h. The net EPV tendency differences are dominated by changes in longwave radiation (Fig. 11b). Areas above where vertical cloud gradients are locally weaker correspond to areas where EPV destruction is locally greater, because of reductions in cloud-top longwave radiative cooling.

It is apparent from Fig. 11 that vortex intensification is slower with shortwave radiation because of the reduction in cloud in the vortex core, which reduces EPV generation owing to longwave radiation from cloud-top cooling. To explore the reason for the reduction in clouds, we examine the radiative heating rate differences between E4 and E5 in Fig. 12. Heating results in reduced cloud, which results in less infrared radiative cooling (Fig. 12a). Shortwave radiative heating exhibits a broader pattern than longwave radiation, and does not correlate
with changes in clouds as for longwave radiation (Fig. 12b). Shortwave heating rates are larger where water vapor decreases rapidly with height. These results indicate that shortwave radiative heating rates are more sensitive to the changes in vertical gradients in water vapor than to clouds. Longwave radiation also acts to cool the layer above the cloud, which is a destabilizing effect that promotes cloud maintenance (e.g., Fu et al. 1995). However, shortwave radiative heating partially offsets the cooling from longwave radiation above the clouds. Over time, the existence of lower cloud concentrations leads to further reductions in longwave cooling and less intensification with respect to E4. However, even if clouds were to completely diminish, we still expect the vortex to intensify given the results of experiment E3.

4. Summary and conclusions

The Arctic is a favorable region for the maintenance of tropopause polar vortices (TPVs) because radiative processes dominate latent heating. Here, five sets of idealized numerical modeling experiments were designed to examine the physical mechanisms leading to TPV intensification under the hypothesis that stronger longwave cooling rates owing to anomalously dry air inside the vortex create relatively larger amounts of EPV near the tropopause.

Results show that anomalously dry air inside the vortex core is an essential aspect of vortex intensification. In particular, longwave radiative cooling is stronger where water vapor decreases more rapidly with height than in the surrounding environment. This effect dominates the tendency from the vortex temperature structure, ozone, and carbon dioxide. Since the tropopause extends downward to relatively higher pressure in the vortex, a vertically thin transition zone is established between dry stratospheric air and moist tropospheric air. This results in a large positive gradient in longwave heating between the upper troposphere and lower stratosphere, creating EPV across the tropopause and strengthening the vortex by ~17% in 5 days. When water vapor is excluded, the vortex still intensifies, however, only by ~5% in 5 days, and the EPV tendency does not resemble observed patterns (Cavallo and Hakim 2010). Nevertheless, the dry case is interesting because the vortex continues to strengthen for ~150 days, primarily because strong initial relative vorticity projects onto the vertical longwave heating gradient, creating EPV in the vortex core.

Maximum intensification is found when clouds contribute to longwave radiation in the absence of solar radiation, characteristic of Arctic winter; in this experiment, the vortex strengthens by ~34% in 5 days. Enhanced cloud-top cooling in the upper troposphere results in strong EPV creation, similar to that found by Cavallo and Hakim (2009). Including shortwave radiation, as in Arctic summer, slightly reduces vortex intensification over 5 days to ~19%. While it is clear that these radiative mechanisms provide suitable conditions for vortex intensification in the short term, the longer-term limits of intensification would be interesting to examine in future studies given that TPVs can be observed to exhibit lifetimes of weeks to months.

Decreases in Arctic sea ice will likely result in substantial increases of sensible and latent heat fluxes into the atmosphere, particularly during the autumn and winter months when air–sea gradients are relatively large (e.g., Serreze et al. 2007; Screen and Simmonds 2010). These changes may alter the radiative effects on TPVs, which could in turn affect the low-level atmospheric circulation. Such large-scale changes have already been seen in global climate modeling (GCM) studies (e.g., Deser et al. 2010; Higgins and Cassano 2009; Screen et al. 2012, 2013), but the connection, if any, to TPVs remains unclear.

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APPENDIX A

Vortex Specification

The initial vortex is initialized using a collection of software called 3DVPAS (see Nolan and Montgomery 2002; Nolan and Grasso 2003; Hodys and Nolan 2007; Nolan et al. 2007). 3DVPAS is based on the dry, anelastic momentum equations in cylindrical coordinates and allows for the creation of a basic-state axisymmetric vortex in gradient wind and hydrostatic balance based upon observed data. Tangential velocity, moisture, and temperature data from Cavallo and Hakim (2010) are used to create the initial vortex. All calculations in 3DVPAS are made on a radius–height (r–z) plane with 0 ≤ r ≤ 1452 km, 0 ≤ z ≤ 22 km, 121r × 61z grid points, and a constant horizontal eddy viscosity of ν = 40 m² s⁻¹.

A1 All initialization computations were performed using MATLAB 6.5.0.
Radial profiles for use in 3DVPAS are obtained here by interpolating the winter (December–February) vortex structure onto the \( r-z \) grid. The tangential wind is filtered using a five-point smoother and is mapped onto a radial grid one level at a time by applying a first-order (fourth-order) polynomial data fit in the stratosphere (troposphere) and onto the vertical grid independently by piecewise cubic spline interpolation on each radial grid point. Zero tangential wind is imposed at both the vortex core and at the outer boundary of the radial grid. A radial weighting function, \( w(r) \), smooths sharp gradients in the tangential wind inside the vortex core:

\[
w(r) = \begin{cases} 
\cos[-\pi(r-r_{\text{in}})/2r_{\text{out}}], & r \geq r_{\text{emax}} \\
\sin(\pi r/2r_{\text{in}})^\alpha, & r < r_{\text{emax}} 
\end{cases}
\tag{A1}
\]

where \( r_{\text{emax}} \) is the radius of maximum wind, \( \alpha = 0.03 \), and \( r_{\text{out}} (r_{\text{in}}) \) are the radii from the radius of maximum wind to the domain edge (radius of maximum wind to the vortex core). Similarly, a vertical weighting function, \( w(z) \), which damps velocities near the top of the model, is also applied to the tangential wind:

\[
w(z) = \begin{cases} 
\cos[\pi(z-z_{\text{damp}})/2L], & z \geq z_{\text{damp}} \\
1, & \text{otherwise}
\end{cases}
\tag{A2}
\]

where \( L = z_{\text{top}} - z_{\text{damp}} \), and \( z_{\text{top}} \) and \( z_{\text{damp}} \) are the heights at the top of the model and at the level 30 grid points from the top of the model, respectively. The mean vortex temperature sounding is taken from the composite data at the column corresponding to the maximum tangential wind. The temperature data is initially smoothed using a five-point moving average in the stratosphere beginning from the tropopause. Then it is linearly interpolated onto the vertical grid and extrapolated outside the range of the composite data of \( \sim 13.5 \) km.

The composite data from Cavallo and Hakim (2010) indicate that the troposphere is nearly saturated in the vicinity of TPVs. As the net effect of longwave radiation is to cool the atmosphere, it may act to quickly cool the atmosphere to the saturation point, especially in the absence of shortwave radiation. This is an important consideration when choosing an initial water vapor field since initializing with subsaturated air may create a transitional period until saturation when cloud formation will subsequently occur. Here, we choose to begin the experiments using water vapor with relative humidity values equal to 100% with respect to liquid water in the troposphere, as the goal is not to understand the transition state before cloud formation, but rather to understand how radiation and latent heating affect vortex intensity. In the stratosphere, the water vapor mixing ratio becomes nearly constant with height, and here a single, fixed water vapor mixing ratio of \( 4.56 \times 10^{-6} \) kg kg\(^{-1} \) from the composite water vapor mixing ratio sounding is used in the stratosphere. A five-point smoother is used to reduce the potential for spurious, unphysical effects that can sometimes result from sharp water vapor gradients at the tropopause from these specifications.

**APPENDIX B**

**Initialization**

WRF is run for 40 min with an output interval equal to the model time step (2 min). The fields are averaged in time, and the vortex field is balanced by removing asymmetries and setting radial and vertical velocities to zero; the process is iterated 10 times. After this procedure is performed, a final check on the vortex steadiness is provided by the mass conservation equation in WRF, \( \partial \mu / \partial t + (V \cdot \nabla \mu) = 0 \). Here \( \mu \) represents the mass per unit area within the column in the model domain at \((x, y)\), \( V = (\partial / \partial x, \partial / \partial y, \partial / \partial \eta) \) is the gradient operator, \( \eta \) is the vertical coordinate, and \( V = (u, v, w) \) is the three-dimensional velocity vector (Skamarock et al. 2005). A 30-min test simulation is performed with a model time step and output interval of 120 s, and no model physics. To examine the initial imbalance of the vortex in the WRF simulations, the mean absolute mass tendency is computed as

\[
\overline{|\mu|_t} = \frac{1}{N} \sum_i \sum_j |{\partial \mu \over \partial t}|
\tag{B1}
\]
where $i$ and $j$ are $i$th and $j$th horizontal grid points and $N = i \times j$. $\Delta \mu / \Delta t$ is calculated using centered differences from model output using each time step and averaged over the domain. Figure B1 shows that the mean absolute mass tendency decreases rapidly within the first few time steps and is relatively small thereafter.

REFERENCES


