<u>Synoptic Meteorology II: Self-Development in the IPV Framework</u>

Readings: Section 5.3.6 of *Midlatitude Synoptic Meteorology*.

Introduction

To date, we have developed several fundamental insights from an IPV perspective. These include the three-dimensional structures of upper-tropospheric IPV and surface potential-temperature anomalies, the impacts of diabatic heating and friction on IPV, and how synoptic-scale upper-tropospheric IPV and surface potential-temperature anomalies move. These insights can be used to recast the concept of cyclone self-development in such a way to better describe the cyclone's development and evolution.

There are at least three main advantages to doing so:

- The IPV perspective does not explicitly require information about vertical motion.
- The IPV perspective better describes the concept of the "phase locking" of the upper- and lower-tropospheric trough/ridge patterns.
- The IPV perspective provides us a more complete view of cyclone occlusion.

Before we proceed, it is useful to note that our initial development of the IPV perspective on self-development neglects diabatic heating and friction. We will consider the role that these processes play at a later point in this lecture. Likewise, note that we are considering surface cyclone development on level ground.

Step 1: Surface Cyclogenesis

Consider a cyclonic upper-tropospheric IPV anomaly superimposed on a background meridional IPV gradient, with IPV increasing toward the poles, and in an environment of westerly vertical wind shear. We assume that this anomaly is of relatively short wavelength such that it moves from west to east at a velocity slower than that of the upper-tropospheric westerly flow.

Consider what happens as this anomaly encounters a lower-tropospheric baroclinic zone with potential temperature increasing toward the Equator across the baroclinic zone. The positive upper-tropospheric IPV anomaly is associated with a weak cyclonic surface circulation, the strength of which depends on the upper-tropospheric cyclonic IPV anomaly's strength and scale. The weak cyclonic surface circulation fosters a dipole of temperature advection, with warm-air advection to the east and cold-air advection to the west, that creates a warm surface temperature anomaly to the east and a cold surface temperature anomaly to the west. Since we are on level ground, these surface temperature anomalies can equivalently be viewed as surface potential-temperature anomalies.

Recall that positive surface potential-temperature anomalies are akin to cyclonic upper-tropospheric IPV anomalies and negative surface potential-temperature anomalies are akin to anticyclonic upper-tropospheric IPV anomalies. As a result, the surface potential-temperature anomalies are also surface rotation anomalies, with cyclonic rotation developing to the east and anticyclonic rotation developing to the west. (As an aside, note that if we did not have an initial baroclinic zone, none of this would be able to take place and thus surface cyclogenesis could not occur!)

This development stage is conceptualized in Fig. 1.

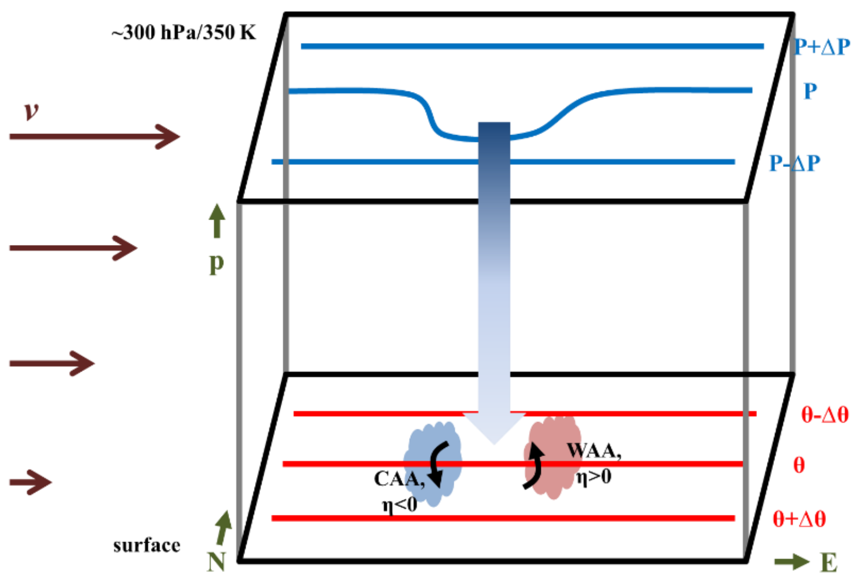

Figure 1. Schematic of a positive upper-tropospheric IPV maximum inducing (downward arrow) a surface cyclonic circulation (black arrows) and warm (red cloud) and cold (blue cloud) potential-temperature advection along a surface baroclinic zone (red contours). Isopleths of IPV are depicted in blue contours, whereas the horizontal wind field is depicted by dark red arrows.

Step 2: Surface Development

The warm and cold surface potential-temperature anomalies are associated with weak upper-tropospheric cyclonic and anticyclonic circulations, respectively, the strength of which are directly proportional to the strengths and length scales of the surface potential-temperature anomalies. Since the surface potential-temperature anomalies are east of the upper-tropospheric

IPV anomalies, their associated circulations aloft are also east of the upper-tropospheric IPV anomalies.

These associated upper-tropospheric circulations impart a northerly flow into the core of the cyclonic upper-tropospheric IPV anomaly. Because the background IPV increases to the north, this northerly flow is associated with positive IPV advection that intensifies the anomaly. Since this advection is in the core of the cyclonic upper-tropospheric IPV anomaly and not to its west, this positive IPV advection also helps reduce the upper-tropospheric anomaly's tendency to retrogress westward. This allows the cyclonic upper-tropospheric IPV anomaly to move eastward more rapidly than it did before, albeit still at a slower velocity than that of the upper-tropospheric westerly flow.

This development stage is conceptualized in Fig. 2.

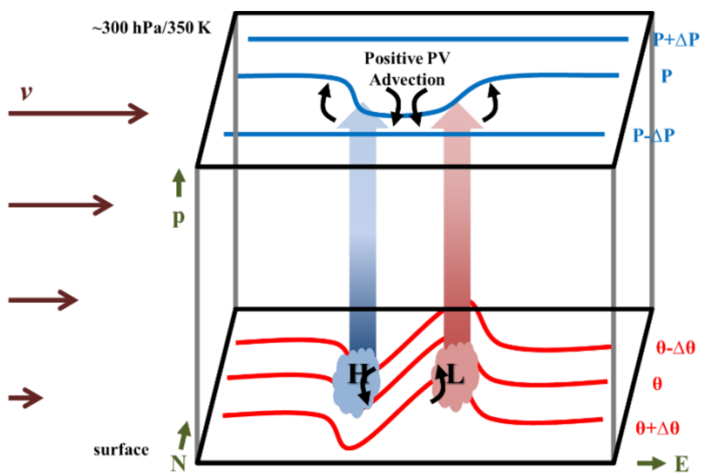

Figure 2. Schematic of warm (red cloud) and cold (blue cloud) surface potential-temperature anomalies inducing (upward arrows) upper-tropospheric cyclonic and anticyclonic circulations (black arrows on top surface), respectively, and positive IPV advection across the IPV gradient (blue contours). Isentropes are depicted in red, whereas the horizontal wind field is depicted by dark red arrows.

In turn, strengthening the cyclonic upper-tropospheric IPV anomaly intensifies its associated cyclonic flow at the surface. This then intensifies the lower-tropospheric dipole of potential-temperature advection, thereby intensifying the warm and cold surface potential-temperature anomalies and thus the surface cyclone and anticyclone, respectively. It also assists in reducing the tendency for the surface pattern to move eastward (toward the greatest warm-air advection, given the direct relationship between warm surface potential-temperature anomalies and cyclonic rotation noted above) by enhancing the warm and cold surface potential-temperature advection to the west of where the anomalies themselves would focus it.

Thus, the natural tendency of the upper-tropospheric IPV and surface potential-temperature anomalies to propagate in opposite directions from one another is reduced by the circulations associated with each other! This allows for upper- and lower-tropospheric anomalies of like sense (e.g., positive/warm and negative/cold) to remain more optimally located with respect to each other, with a slight westward tilt with increasing height, during cyclone development. This is the IPV-based interpretation of "phase locking." The vertical tilt to the west with increasing height *must* be present for development to occur; were there to be no tilt or a tilt to the east with increasing height, development would halt or reverse.

In turn, intensifying the warm and cold surface potential-temperature anomalies then intensifies their associated upper-tropospheric circulations. Their influences on the upper-tropospheric pattern are as described above, just stronger. These then feed back to the surface, which then feed back to the upper troposphere, and so on. Thus, just as we saw with the Pettersen-Sutcliffe development framework, a feedback loop that characterizes self-development is established!

Step 3: Development to Maturity

From this point onward, the basic interpretation of self-developing in the IPV framework is like that from the Pettersen-Sutcliffe development equation until cyclone occlusion. Over time, the potential-temperature advection by the surface cyclone deforms the isotherms sufficiently to reduce the magnitude of warm air advection found to the east and, eventually, north of the cyclone. This deflects the surface cyclone's motion more northward and less eastward.

Aloft, IPV advection by the circulation associated with the surface warm potential-temperature anomaly gradually causes the positive upper-tropospheric IPV anomaly to acquire a negative horizontal tilt. Furthermore, it gradually deforms the IPV isopleths, reducing the magnitude of positive IPV advection to the west and, eventually, south of the cyclone. This further reduces the tendency of the anomaly to retrogress westward, allowing it to move eastward at yet still a faster velocity – reducing the cyclone's vertical tilt!

Step 4: From Maturity to Occlusion

The slowed eastward progression of the surface pattern and accelerated eastward progression of the upper-tropospheric pattern (relative to their natural tendencies) allow for the cyclonic upper-tropospheric IPV anomaly to "catch up" to the surface cyclone and reduce the cyclone's vertical tilt. Eventually, the vertical tilt becomes zero, and the cyclone becomes vertically stacked. The initial vertical stacking of the upper- and lower-tropospheric anomalies represents the maturity stage of the surface cyclone's lifecycle. No longer can one anomaly mutually amplify the other by advections driven by their respective associated circulations.

At and after this point, the vertically stacked system begins to fill as mass is laterally transported into the cyclone volume throughout the troposphere; there is no longer any means of evacuating it away from the cyclone at any level. Friction directly weakens the surface cyclone and, through thermal vorticity arguments, creates a cold surface potential-temperature anomaly that fosters the weakening of the cyclonic upper-tropospheric IPV anomaly.

Influences of Diabatic Processes

It is useful to consider how diabatic heating can influence the decay and development processes in the IPV self-development framework. East of the surface cyclone, relatively warm poleward-moving air parcels ascend as they encounter the baroclinic zone along which the surface cyclone first developed. If sufficient lift and moisture are present, clouds and precipitation may form as water vapor condenses. Indeed, north – from northwest to northeast – of a surface warm front is a favored location for cloud and precipitation formation with midlatitude extratropical cyclones, as we demonstrated earlier in the contexts of conveyor belts and isentropic analysis.

Diabatic warming associated with latent heat release due to condensation is thus found in the lower to middle troposphere near and to the northeast of the surface cyclone. Previously, we demonstrated that this vertical configuration of diabatic warming increases IPV below the diabatic-warming maximum. Therefore, diabatic warming maximized in the midtroposphere intensifies the surface cyclone! In other words, midtropospheric diabatic warming allows surface cyclogenesis to proceed more rapidly and/or to be more intense than if it were not present – just as in the Pettersen-Sutcliffe development framework!

We can also consider the impact of diabatic warming on upper-tropospheric IPV. As clouds and precipitation wrap westward around the surface cyclone, the associated diabatic warming reduces the upper-tropospheric IPV. This occurs first on the northeastern flank of the positive upper-tropospheric IPV anomaly and advances westward with time. This diabatic influence is augmented by the adiabatic advection of low IPV air westward along the northern flank of the cyclonic upper-tropospheric IPV anomaly by the circulation associated with the warm surface potential-temperature anomaly. Thus, due to both adiabatic and diabatic processes, the positive upper-tropospheric IPV anomaly takes on an increasingly large negative tilt and can cut off from the synoptic-scale flow as the cyclone occludes!

The horizontal structure of a positive upper-tropospheric IPV anomaly that is in the process of cutting off is depicted in Fig. 3. An overview of the impact of diabatic heating on cyclone occlusion and cut-off development is depicted in Fig. 4. This lifecycle is known as the LC2 synoptic cyclone lifecycle (Thorncroft et al. 1993; *Quart. J. Royal Meteor. Soc.*), and the cyclonic upper-tropospheric IPV anomaly's evolution is known as cyclonic wavebreaking (so named because the upper-tropospheric IPV isopleths turn cyclonically like a wave in the ocean).

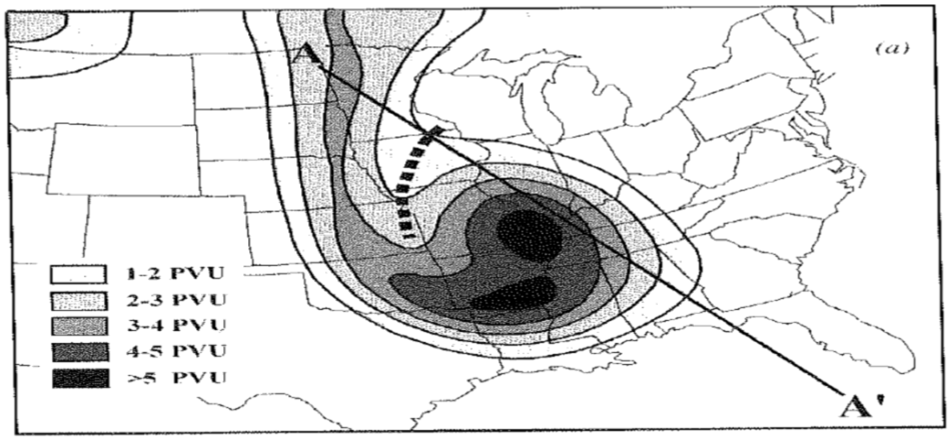

Figure 3. Isopleths of potential vorticity (contoured and shaded; units: PVU) on a generic upper-tropospheric isobaric surface. Reproduced from *Mid-Latitude Atmospheric Dynamics* by J. Martin, their Fig. 9.15a.

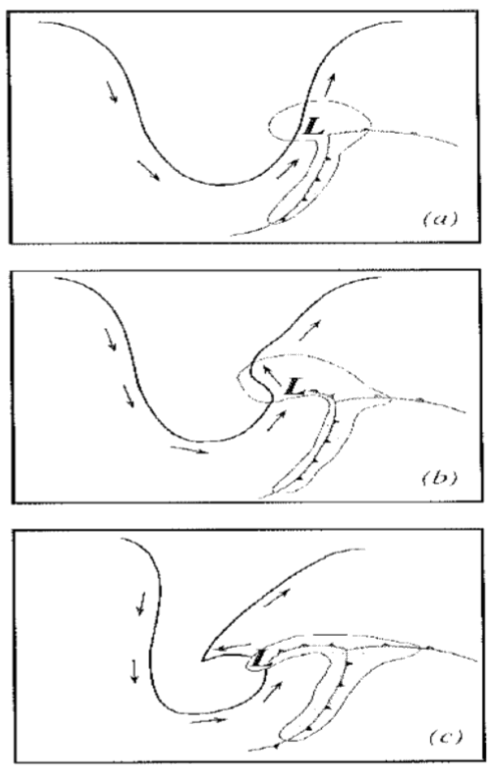

Figure 4. Schematic illustrating the erosion of the northeastern flank of an upper-tropospheric IPV anomaly (solid black contour) by negative IPV advection (arrows) and diabatic heating (cloud surrounding the surface cyclone system). Reproduced from *Mid-Latitude Atmospheric Dynamics* by J. Martin, their Fig. 9.16.

The Development of Upper-Tropospheric Troughs

Whether in the Pettersen-Sutcliffe or IPV self-development frameworks, we have assumed the presence of a pre-existing upper-tropospheric trough superimposed on a lower-tropospheric baroclinic zone. While self-development illustrates how upper-tropospheric troughs can amplify and occlude, it does not provide a framework to explain their initial formation. Instead, other arguments are needed. These arguments are typically tied to *barotropic or baroclinic instability*, with the latter manifest generally in terms of the Eady and/or Charney model. We will not cover either instability in this class; however; you may encounter these and related concepts in Dynamics II and/or subsequent graduate-level courses.