Synoptic Meteorology II: Self-Development in the IPV Framework

5-7 May 2015

Readings: Section 5.3.6 of *Midlatitude Synoptic Meteorology*.

Introduction

In this and other recent lectures, we have developed a number of fundamental insights from the perspective of isentropic potential vorticity. These include, but are not necessarily limited to, a) the structures of upper tropospheric isentropic potential vorticity and surface potential temperature anomalies, b) the "action at a distance" principle, c) the impacts of diabatic heating and friction upon isentropic potential vorticity, and d) the motion of synoptic-scale upper tropospheric isentropic potential vorticity and surface potential temperature anomalies. These insights can be used to recast the concept of "self-development" in such a way so as to more eloquently describe surface cyclone development and evolution.

There are at least three main advantages to doing so:

- The isentropic potential vorticity perspective on "self-development" does not require information about vertical motion.

- The isentropic potential vorticity perspective on "self-development" can be invoked to better describe the concept of the "phase locking" of the upper and lower tropospheric trough/ridge patterns, at least in a qualitative sense.

- The isentropic potential vorticity perspective on "self-development" provides us with perhaps a more complete view of the cyclone occlusion process.

Before we proceed, it is useful to note that our initial development of the isentropic potential vorticity perspective upon "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

Let us consider a positive upper tropospheric isentropic potential vorticity anomaly that is superimposed upon a background meridional gradient in isentropic potential vorticity with isentropic potential vorticity increasing toward the poles. This anomaly is located within an environment characterized by westerly vertical wind shear. We assume that this anomaly is of moderately 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 such as an old frontal boundary. Note that we presume that potential temperature increases toward the equator across this baroclinic zone. The positive upper tropospheric isentropic potential vorticity anomaly, by the "action at a distance" principle, induces a weak cyclonic circulation at the surface. The precise strength of this induced circulation depends upon the strength and scale of the positive upper tropospheric isentropic potential vorticity anomaly, as we discussed in our lecture entitled "Potential Vorticity Inversion and Anomaly Structure."

The induced cyclonic circulation at the surface results in a dipole of temperature advection, with warm air advection to the north and east and cold air advection to the south and west. This creates a warm surface temperature anomaly to the east and a cold surface temperature anomaly to the west of the induced cyclonic circulation. Since we are on level ground, these surface temperature anomalies can equivalently be viewed as surface potential temperature anomalies.

In an earlier lecture, we demonstrated that warm surface potential temperature anomalies are analogous to positive upper tropospheric isentropic potential vorticity anomalies and that cold surface potential temperature anomalies are analogous to negative upper tropospheric isentropic potential vorticity anomalies. As a result, surface cyclogenesis occurs to the east and surface anticyclogenesis occurs to the west of the positive upper tropospheric isentropic potential vorticity anomaly. Note that without the presence of a baroclinic zone, surface cyclogenesis and anticyclogenesis – and, as we well demonstrate shortly, further development – cannot occur!

The development of the warm and cold surface potential temperature anomalies is conceptualized in Figure 1 below.
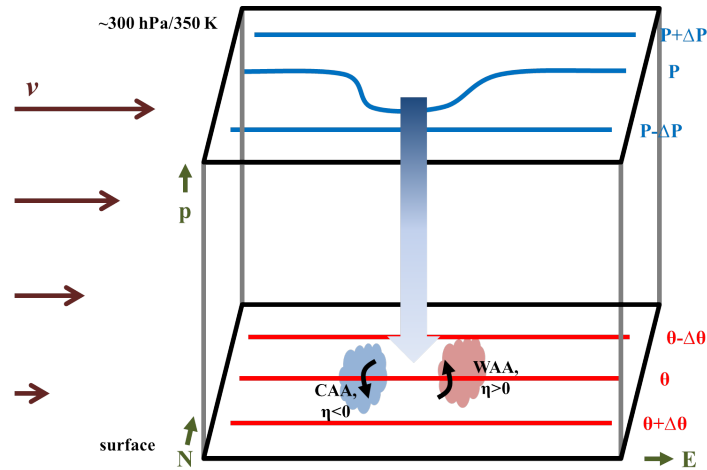
Figure 1. Schematic of a positive upper tropospheric isentropic potential vorticity maximum inducing (downward arrow) a surface cyclonic circulation (black arrows) and warm (red cloud) and cold (blue cloud) potential temperature advection along a surface baroclinic zone (red contours). Isopleths of isentropic potential vorticity are depicted in blue contours, while the vertical structure of the horizontal velocity field is depicted by red arrows.

Step 2: Surface Development

The warm and cold surface potential temperature anomalies induce cyclonic and anticyclonic circulations, respectively, in the upper troposphere. Like their upper tropospheric counterparts, stronger, larger (in scale) surface potential temperature anomalies induce stronger circulations aloft. Owing to the warm surface potential temperature anomaly being located east of the positive upper tropospheric isentropic potential vorticity anomaly, the induced cyclonic circulation aloft is located east of the center of the positive upper tropospheric isentropic potential vorticity anomaly. Likewise, the induced anticyclonic circulation aloft is located west of the center of the positive upper tropospheric isentropic potential vorticity anomaly.

These induced circulations impart a northerly flow into the heart of the positive upper tropospheric isentropic potential vorticity anomaly. This northerly flow is associated with positive isentropic potential vorticity advection that intensifies the anomaly and helps to reduce its tendency to retrogress westward. This allows the positive upper tropospheric isentropic potential vorticity anomaly to move eastward more rapidly than before, albeit still at a slower velocity than that of the upper tropospheric westerly flow.

The impact of the warm and cold surface potential temperature anomalies upon the positive upper tropospheric isentropic potential vorticity anomaly is conceptualized in Figure 2 below.
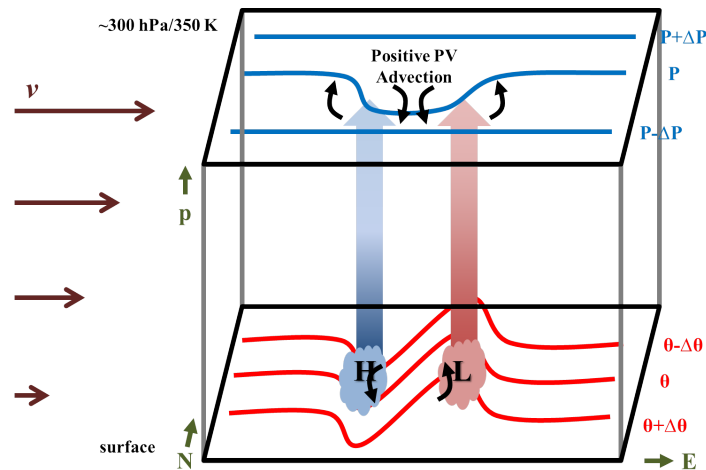
Figure 2. Schematic of warm (red cloud) and cold (blue cloud) surface potential temperature anomalies inducing (upward arrows) upper tropospheric cyclonic and anticyclonic circulations (black arrows on top surface), respectively, and positive isentropic potential vorticity advection across the gradient of isentropic potential vorticity (blue contours). Isentropes are depicted in red contours, while the vertical structure of the horizontal velocity field is depicted by red arrows.

Strengthening the positive upper tropospheric isentropic potential vorticity anomaly acts to intensify the cyclonic circulation that it induces at the surface. This intensifies the lower tropospheric dipole of potential temperature advection, thereby intensifying the warm and cold surface potential temperature anomalies and, by analogy, the surface cyclone and anticyclone respectively. It also assists in reducing the tendency for the surface pattern to progress eastward by enhancing the warm and cold surface potential temperature advection to the west of where the anomalies themselves would focus it.

As a consequence, the natural tendency of the upper tropospheric isentropic potential vorticity and surface potential temperature anomalies to propagate in opposite directions from one another has been reduced. 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 – so as to permit development to proceed, whether more efficiently or at all! This is the isentropic potential vorticity-based qualitative interpretation of the idea of "phase locking." Note that the vertical tilt to the west with increasing height **must** be present in order for development to occur; were there to be no tilt or a tilt to the east with increasing height, development would halt or reverse altogether.

Intensifying the warm and cold surface potential temperature anomalies thus intensifies the circulations that each induces in the upper troposphere. Its influence upon the upper tropospheric

pattern is as above, which then feeds back to the surface, and so on. Thus, as we saw from the quasi-geostrophic perspective on self-development, a feedback loop is established!

Step 3: Development to Maturity

From this point, the basic interpretation is similar to that afforded by the quasi-geostrophic perspective on self-development up until the point of cyclone occlusion. Gradually, thermal advection by the surface cyclone deforms the isotherms sufficiently so as to reduce the magnitude of warm air advection found to the east and, eventually, north of the cyclone. This also slows its eastward propagation.

Aloft, isentropic potential vorticity advection induced by the surface warm potential temperature anomaly gradually causes the positive upper tropospheric isentropic potential vorticity anomaly to acquire a negative horizontal tilt. Furthermore, it gradually deforms the contours of constant isentropic potential vorticity, reducing the magnitude of positive isentropic potential vorticity 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.

Step 4: From Maturity to Occlusion

The slowed eastward progression of the surface pattern and more rapid eastward progression of the upper tropospheric pattern allow the positive upper tropospheric isentropic potential vorticity anomaly to "catch up" to the surface cyclone. Eventually, the upper tropospheric and surface anomalies become vertically stacked. The initial vertical superposition of these two anomalies represents the mature, or most intense stage of the surface cyclone's lifecycle. No longer can one anomaly mutually amplify the other by advections associated with induced circulations.

At and after this point, the vertically-stacked system begins to fill. In other words, mass is laterally transported into the cyclone volume throughout the troposphere; there is no longer any means of evacuating it away from the cyclone in the upper troposphere. Friction acts to dissipate the surface cyclone directly. Furthermore, through thermal vorticity arguments, it creates a cold surface potential temperature anomaly that induces a negative isentropic potential vorticity anomaly in the upper troposphere, thereby weakening the positive upper tropospheric potential vorticity anomaly with time.

The Impact of Diabatic Processes upon Development and Occlusion

It is now useful to consider how diabatic heating can influence both the decay and development process in the self-development framework. To the east of the surface cyclone, relatively warm poleward-moving air parcels ascend as they encounter the baroclinic zone along which the surface cyclone developed. Presuming that 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 mid-latitude extratropical cyclones, as we demonstrated earlier this semester in the contexts of isentropic analysis and conveyor belts.

Diabatic warming associated with latent heat release due to condensation is thus found in the lower to middle troposphere near the surface cyclone and to its north and east. Previously, we demonstrated how this configuration of diabatic warming influences the vertical distribution of isentropic potential vorticity: namely, it increases isentropic potential vorticity below and decreases isentropic potential vorticity above the diabatic warming maximum. In the lower troposphere, therefore, the effect of diabatic warming is to intensify the surface cyclone. Thus, as from the quasi-geostrophic perspective, we state that diabatic warming allows surface cyclogenesis to proceed more rapidly and/or to be more intense than if it were not present!

We can also consider the impact of diabatic warming upon isentropic potential vorticity in the upper troposphere. As the aforementioned clouds and precipitation wrap westward around the surface cyclone, the associated diabatic warming reduces the upper tropospheric isentropic potential vorticity. This occurs first on the northeastern flank of the positive upper tropospheric isentropic potential vorticity anomaly and, with time, works westward.

Likewise, from an adiabatic perspective, the induced cyclonic flow aloft resulting from the warm surface potential temperature anomaly advects low isentropic potential vorticity air westward along the northern flank of the positive upper tropospheric isentropic potential vorticity anomaly. In the context of both adiabatic and diabatic processes, the positive upper tropospheric isentropic potential vorticity anomaly takes on an increasingly large negative tilt and, ultimately, can cut off from the synoptic-scale flow, notably as the cyclone occludes. Thus, the isentropic potential vorticity perspective upon self-development provides a more complete qualitative description of the cut-off and occlusion process of mature mid-latitude synoptic-scale cyclones than does the quasi-geostrophic perspective upon self-development.

The horizontal structure of a positive upper tropospheric isentropic potential vorticity anomaly that is in the process of cutting off is depicted in Figure 3. A conceptual overview of the impact of diabatic heating upon cyclone occlusion and cut-off development is depicted in Figure 4.
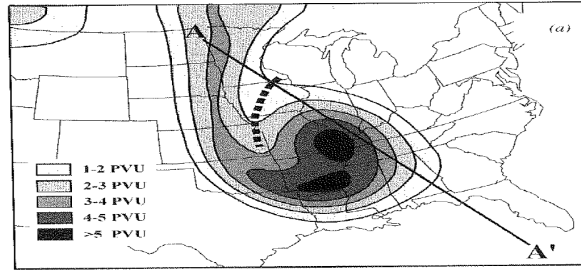
Figure 3. Isopleths of potential vorticity (contoured and shaded; units: PVU) on a generic upper tropospheric isobaric surface. Reproduced from *Mid-Latitude Atmospheric Dynamics* by J. Martin, their Figure 9.15a.
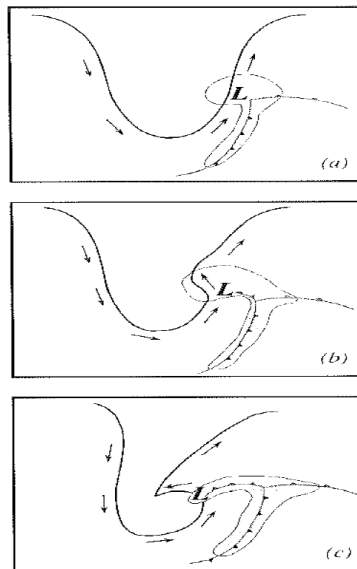

Figure 4. Schematic illustrating the erosion of the northeastern flank of an upper tropospheric isentropic potential vorticity anomaly (solid black contour) by low isentropic potential vorticity advection (arrows) and diabatic heating (cloud surrounding the surface cyclone system). Reproduced from *Mid-Latitude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

The Development of Upper Tropospheric Troughs

When we covered the quasi-geostrophic and isentropic potential vorticity-based versions of the "self-development" framework, we assumed that there was a pre-existing upper tropospheric trough. While the "self-development" framework provides illustrative information about how upper tropospheric troughs can amplify and occlude, it does not provide a framework that helps explain the initial formation of an upper tropospheric trough. Instead, other arguments are

necessary. Oftentimes, such arguments are tied to the concepts of *barotropic instability* or *baroclinic instability*, with the latter manifest generally in terms of the Eady and/or Charney model. We will not cover either barotropic instability or baroclinic instability in this class, however; instead, you will encounter these and related concepts in Dynamics II and/or subsequent graduate-level courses, should you continue onward to graduate school.