National Weather Service

Gaylord, Michigan

The WFO Gaylord Science Corner

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Operational Topics

Potential Vorticity: A Different Perspective

Among the myriad of different ways National Weather Service meteorologists have of slicing and dicing observed and model data, the use of potential vorticity and related conceptual models to describe the atmosphere is relatively new to the operational realm, with these data sets only recently becoming available. But viewing various potential vorticity fields can provided forecasters different insights into upper level features such as the tropopause and jet streaks, processes involving latent heat release and induced circulations from potential vorticity anomalies in the lower levels of the atmosphere, and a new perspective on cyclogenesis resulting from the coupling of upper and lower level potential vorticity anomalies.

This article is intended to serve as an introduction to potential vorticity concepts. The first part of the article will lay out some background information on potential vorticity and related conceptual models. Then we'll look at some operational applications of the potential vorticity approach.

Background

So what exactly is potential vorticity, and what kinds of properties does it have? Potential vorticity, or PV for short, is simply the absolute vorticity on an isentropic (constant potential temperature) surface, multiplied by the static stability. For those who are more mathematically inclined, the equation for PV can be written as the following:

$$PV = -g\left(\zeta_\theta + f\right)\left(\frac{\partial\theta}{\partial p}\right)$$

1 **2**

where term 1 is the absolute vorticity (relative vorticity plus vorticity due to the Earth's rotation, denoted by **f**), and term 2 represents static stability, defined as the change in potential temperature with height. Since potential temperature increases as one goes up in the atmosphere (except in rare cases), the more rapid the increase in potential temperature over a given vertical depth, the more stable that layer of atmosphere is. Based on this definition, we can see that higher values of PV will be associated with higher values of vorticity, and/or higher values of stability.

The units of potential vorticity are rather unusual given the combination of variables that make up PV: $m^2\ s^{-1}\ K\ kg^{-1}$. We can shorten this by defining a potential vorticity unit, or PVU, given typical values of vorticity and stability, as:

$$\mathbf{1\ PVU = 10^{-6}\ m^2\ s^{-1}\ K\ kg^{-1}}$$

We'll come back to potential vorticity units in a minute.

One of the important properties of potential vorticity lies in its conservation for flow that is adiabatic (constant potential temperature) and frictionless. As an example, let's take an air parcel that has an initial value of absolute

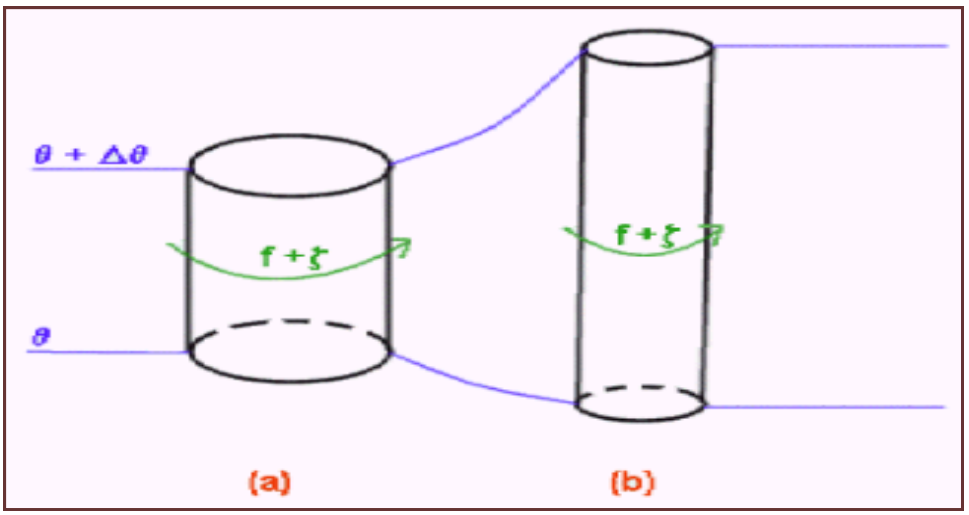

Figure 1. An example of an initial cylindrical air parcel (parcel a on the left), bounded by two isentropic surfaces on the top and bottom, undergoing vertical stretching (parcel b on the right).

vorticity, and is bound on top and bottom by two different isentropic surfaces (recalling that the vertical distance between isentropic surfaces is a measure of stability), as illustrated above as parcel "a" in Figure 1. Thus, parcel "a" has an initial value of potential vorticity based on its' absolute vorticity (denoted by the green arrow), and its' depth (which is a measure of stability). Now let's stretch this parcel vertically, as in parcel "b" in figure 1. Since the flow is adiabatic, the potential temperature at the top and bottom of the parcel does not change, although the distance between these isentropic surfaces does change. This means that we have a change in parcel stability, in this case, stability has decreased. However, since the flow is adiabatic (the potential temperature of the parcel remains constant) the parcel must conserve its' PV as it gets stretched. So going back to the equation that defines PV, which is the product of absolute vorticity and stability, if one variable goes down, the other must go up to compensate. In this case, since stability decreased, the absolute vorticity of the parcel must increase in order for potential vorticity to remain the same. Thus, the parcel must spin faster in order to compensate for the change in stability. Now, if we take this parcel and compress it back to its' original shape, we then have a case where stability is increasing (the isentropic surfaces bounding the parcel are getting closer together), and thus to conserve PV the absolute vorticity must decrease (the parcel must spin slower). Conservation of potential vorticity has important consequences when describing phenomenon such as troughing to the lee of mountain ranges due to flow over the barrier, an example of which is given in Figure 2.

We can get an idea of typical atmospheric values of potential vorticity by looking at a vertical cross section of atmosphere. Figure 3 shows a north-south cross section of average mid latitude potential vorticity and potential temperature. The first thing to note is the uniformly low values of potential vorticity (at or below 1 PVU) found in the troposphere (generally below 8-10km), and the rapid increase in potential vorticity above 10km as one enters the stratosphere (above the red line in Figure 3). The marked change in potential vorticity between the troposphere and the stratosphere is due to the very strong stability in the stratosphere, which contributes to high values of potential vorticity as compared to the more weakly stratified troposphere. The red line in Figure 3 demarcates the dynamic tropopause, a level at which potential vorticity gradients on an isentropic surface are maximized. This level typically ranges from 1.5 to 3 PVU, though operationally the 1.5 or 2 PVU surfaces are used to define the dynamic tropopause. Figure 3 shows a gradually sloping dynamic tropopause from the polar regions to the tropics, though as we shall see the height of the dynamic tropopause can vary greatly.

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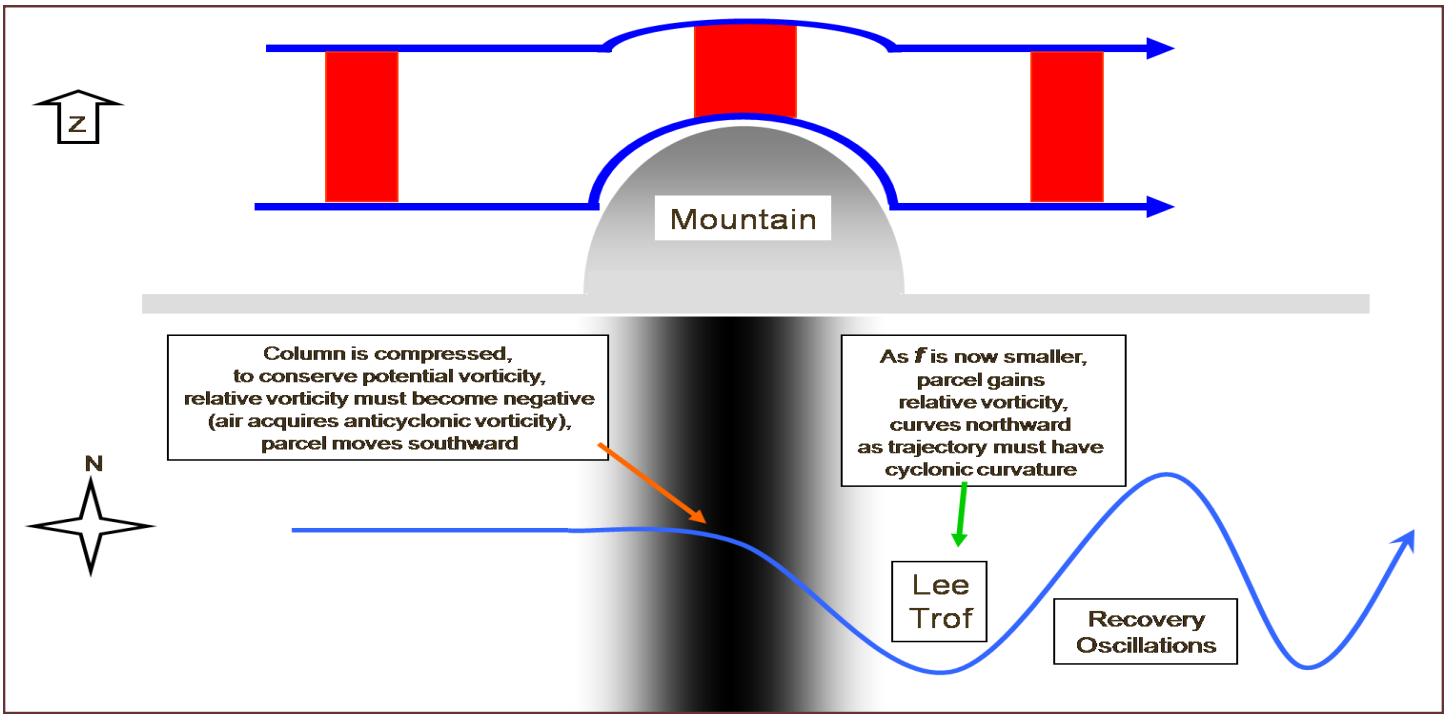

Figure 2. Top: Air parcel in westerly flow undergoing adiabatic compression/expansion due to flow over a topographic barrier. The parcel initially has zero relative vorticity (i.e., its' initial vorticity is only that due to the Earth's rotation, denoted by f. Bottom: Horizontal trajectory of same parcel as a result of potential vorticity conservation. Parcels undergo an anticyclonic trajectory crossing the mountain in order to decrease their vorticity as they are compressed, then undergo a cyclonic trajectory to the lee of the barrier as they increase their vorticity (acquire cyclonic rotation) as they are stretched again downwind of the barrier. Adapted from Holton (1979, **An Introduction to Dynamic Meteorology**) by Dave Baggaley, Prairie and Arctic Storm Prediction Centre, Winnipeg.

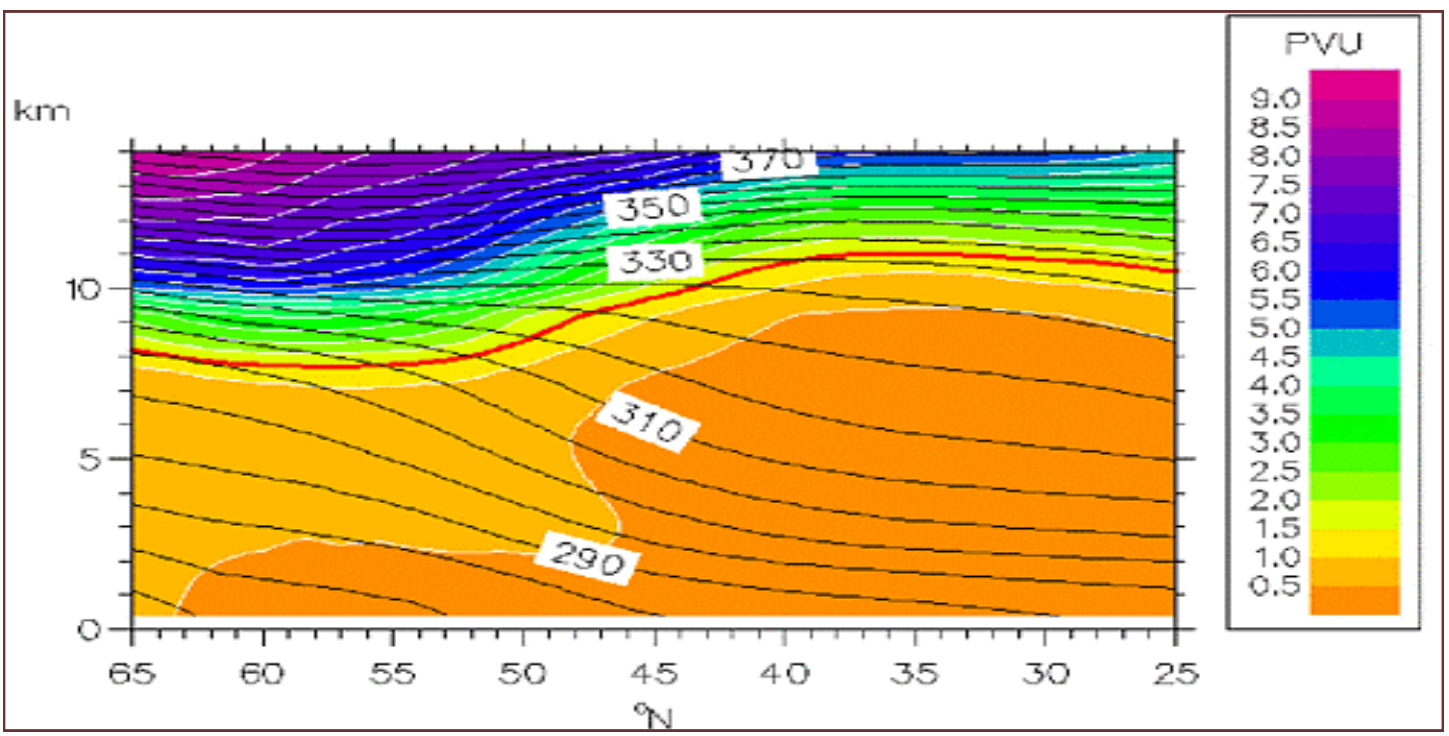

Figure 3. North-south cross section of potential temperature (black contours) and potential vorticity (white contours and color shading), based on a ten year climatology of cold season zonal flow events. The red line denotes the 1.5 PVU surface, also known as the dynamic tropopause. The dynamic tropopause separates low values of potential vorticity in the troposphere from higher values found in the stratosphere.

Since the height of the dynamic tropopause varies, let's take a look at a plan view map of the dynamic tropopause, and compare that with a more traditional 500mb height analysis. We'll use a map of pressure on the 1.5 PVU surface as our proxy for the dynamic tropopause, as shown in figure 4. The figure below shows dynamic tropopause pressure greater than 300mb, with each color band denoting a 50mb increase in pressure. The higher the pressure of the dynamic tropopause, the lower it is in altitude. Notice the values of higher pressure over California, showing the dynamic tropopause down to the 700mb level or so...with an "extension" of lower tropopause height extending back to the east across the northern tier of states and into the Great Lakes. A broad area of lower tropopause pressure/higher tropopause height is found across the central and southeastern U.S....as well as across Alaska and the Yukon Territories.

We can view the relationship between the dynamic tropopause and atmospheric flow pattern in figure 5, which shows the same dynamic tropopause pressure map as in figure 4, but with 500mb height contours overlaid. The broad area of higher tropopause pressure from central/western Canada south into the western U.S. is associated with a mean positively tilted 500mb trough, with the lowest tropopause height over California associated with a strong short wave in the base of the mean trough. The very low dynamic tropopause heights (higher pressure values) north of the Arctic Circle (at the very top of the map over northern Nunavut) are associated with a polar vortex. Higher tropopause heights (lower pressure values) over Alaska and the northeast Pacific ocean are associated with a cutoff high...with higher 500mb heights over the central and southeastern U.S. also associated with lower dynamic tropopause pressures. This is not unexpected...intrusions of stratospheric air farther down into the troposphere are associated with upper level troughs and lows, with the tropopause being higher up beneath upper level ridges and highs.

These centers of higher potential vorticity ($\geq$ 1.5 PVU) are called "potential vorticity anomalies", or "PV anomalies" for short. A distinction can be made between positive PV anomalies (associated with cyclonic disturbances) and negative PV anomalies (associated with anticyclonic circulations/ridges), but from a forecasting perspective we are mainly interested in the positive (cyclonic) anomalies. Positive anomalies at upper levels are often referred to as "dynamic tropopause anomalies", though higher than "typical" values of potential vorticity are not found exclusively at upper levels of the atmosphere, which we'll discuss in a bit.

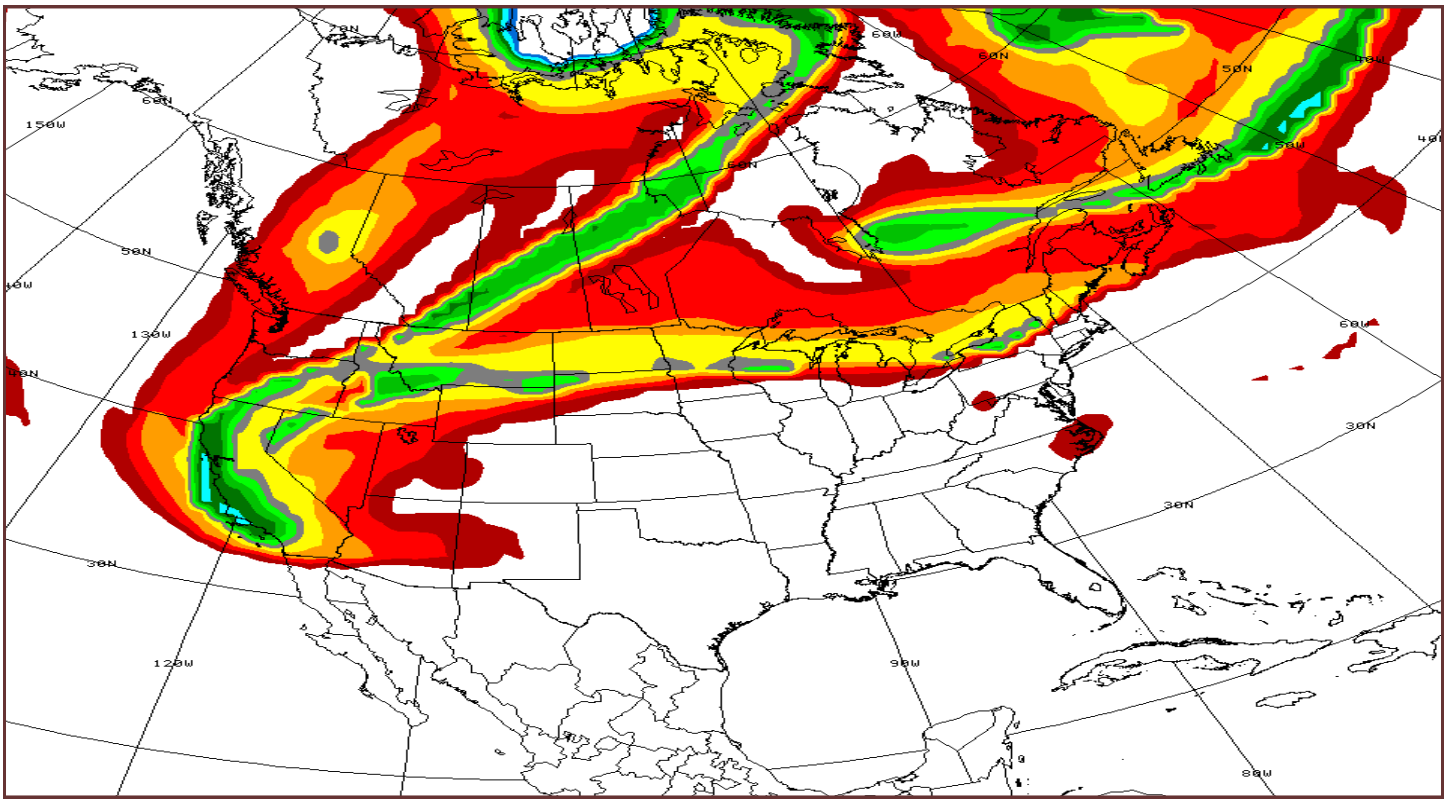

Figure 4. Pressure on the 1.5 PVU surface greater than 300mb. Each color band represents 50mb (i.e., dark red 300-350mb, red 350-400mb, orange 400-450mb, yellow 450-500mb, etc). Areas that are not colored (with the exception of the circular area of high pressure over northern Nunavut at the top of the map) are areas of dynamic tropopause pressure below 300mb (or where the dynamic tropopause is higher in altitude). The white area over northern Nunavut (surrounded by green and blue bands) represents a dynamic tropopause pressure around 800mb).

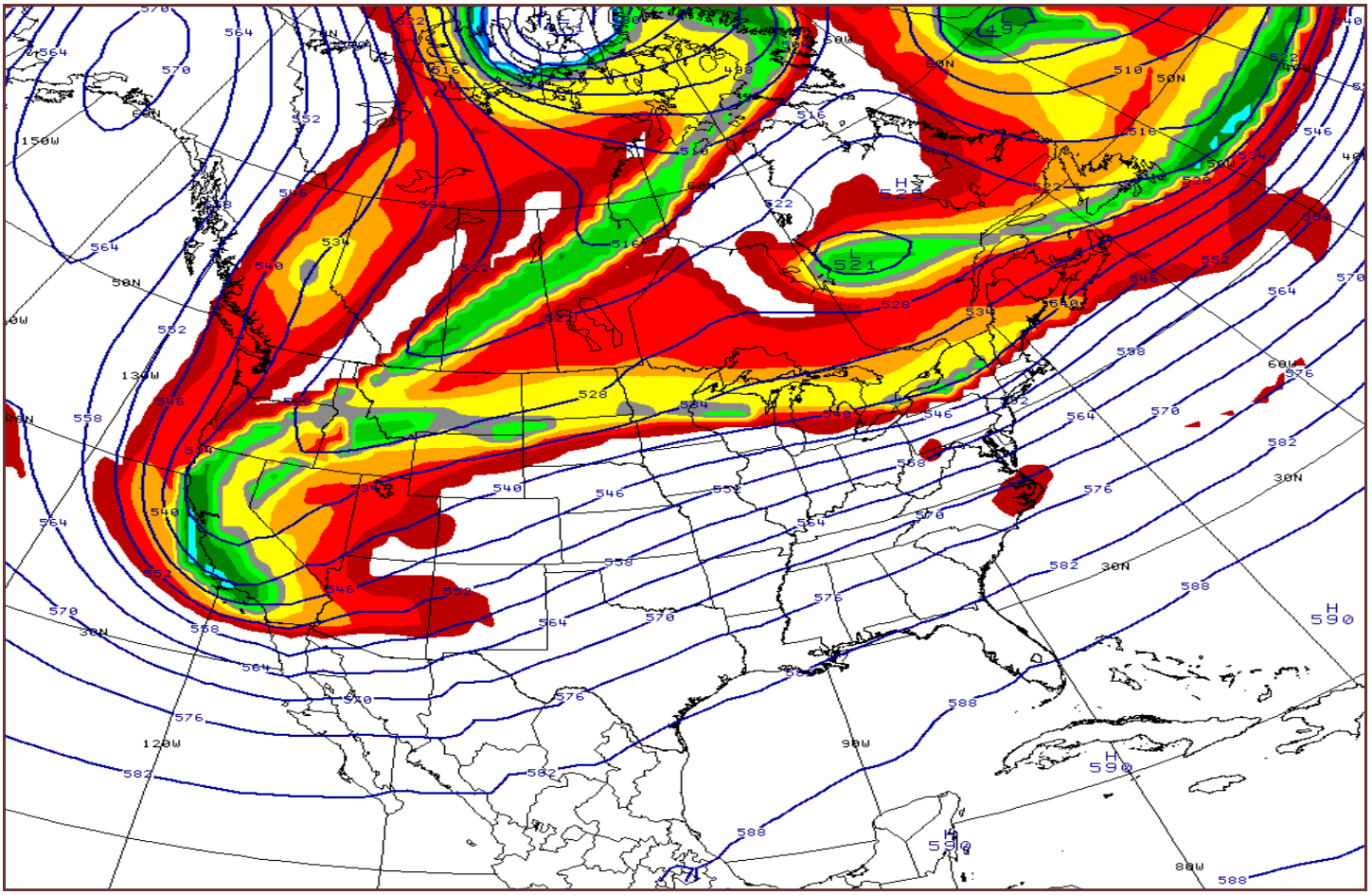

Figure 5. Same as figure 4, except with 500mb heights overlaid (solid contours).

Figure 6 at the right is a vertical cross section taken through a dynamic tropopause anomaly associated with a strong polar vortex (around 484dm at 500mb) over Manitoba province. Note the downward extension of high potential vorticity stratospheric air to the 750mb level.

So we've established that areas of lower dynamic tropopause height are associated with lower geopotential heights...with stronger cyclonic systems having stronger positive PV anomalies (high dynamic tropopause pressure). Another thing to notice in figure 6 is the vertical potential temperature gradient associated with the polar vortex dynamic tropopause anomaly. Notice how the potential temperature contours (isentropes) below the anomaly "bend" upward, while the isentropes above the anomaly bend downward. The resultant tightening of the vertical potential temperature gradient leads to increased stability. So we see both terms of the potential vorticity equation working together...high values of vorticity associated with an upper low, combined with high values of static stability leading to high values of potential vorticity.

This relationship between isentropes and the tropopause anomaly also has interesting implications with regard to vertical motion as the anomaly propagates as shown in figure 7, which is another vertical cross section look at a dynamic tropopause

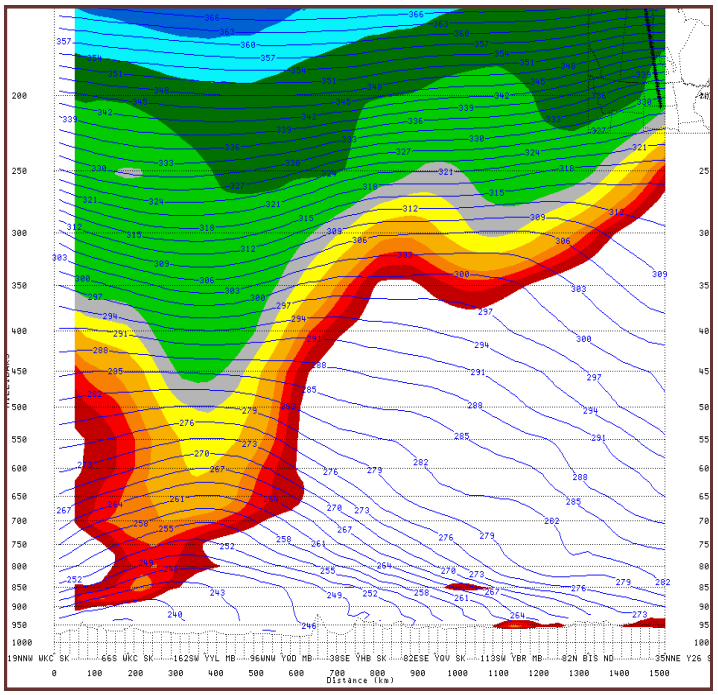

Figure 6. Vertical cross section through a strong polar vortex. Image is potential vorticity greater than 1.5 PVU, contours are potential temperature every 3K.

anomaly. Notice again how the isentropes bend upward toward the lowering dynamic tropopause. Now imagine the tropopause disturbance moving toward the right side of the cross section. As it does, the isentropes out ahead of the disturbance will then rise, or bend upward toward the upper level disturbance, while the isentropes in its wake sink downward. Thus a pattern of rising and sinking motion develops ahead of and behind the travelling dynamic tropopause anomaly, respectively.

Low level potential vorticity anomalies: Thus far, we have discussed exclusively potential vorticity anomalies in the upper levels of the atmosphere. But positive and negative PV anomalies can be found closer to the surface as well. Figure 8 shows a highly idealized cross sectional look at both positive and negative low level PV anomalies. The main thing to take away from figure 8 is that positive (negative) low level PV anomalies are associated with warm (cold) potential temperature anomalies. The top part of figure 8 shows a positive PV anomaly associated with a "bubble" of warm air in the boundary layer, which results in a more stable near surface profile as isentropes bend down toward the surface. The warming induces pressure falls and low level convergence, which increases low level cyclonic vorticity. So a combination of increasing vorticity and increasing stability leads to the development of higher values of potential vorticity. As an aside, it seems backwards to think of stability being associated with low level warming, but notice the spreading out of the isentropes above the surface warm bubble, which is where the stability is actually decreasing. So although is seems confusing at first, the atmosphere is responding as we would expect. The opposite is happening in the bottom diagram in figure 8 with a boundary layer cold "bubble"...the stability is decreasing near the surface (but increasing aloft), and pressure rises lead to divergence and anticyclonic vorticity, and therefore weaker values of potential vorticity. This association between low level warm temperature anomalies and positive potential vorticity has application with regard to looking at cyclogenesis from a PV perspective, which we will look at next.

PV Application: Cyclogenesis

We can start to tie the previous discussions together with a look at cyclogenesis from the perspective of potential vorticity. The distortion of low level isentropes by an upper potential vorticity anomaly discussed in figure 7 has some interesting implications with regard to cyclogenesis, based on the concept of PV conservation mentioned earlier, and recalling the relationship

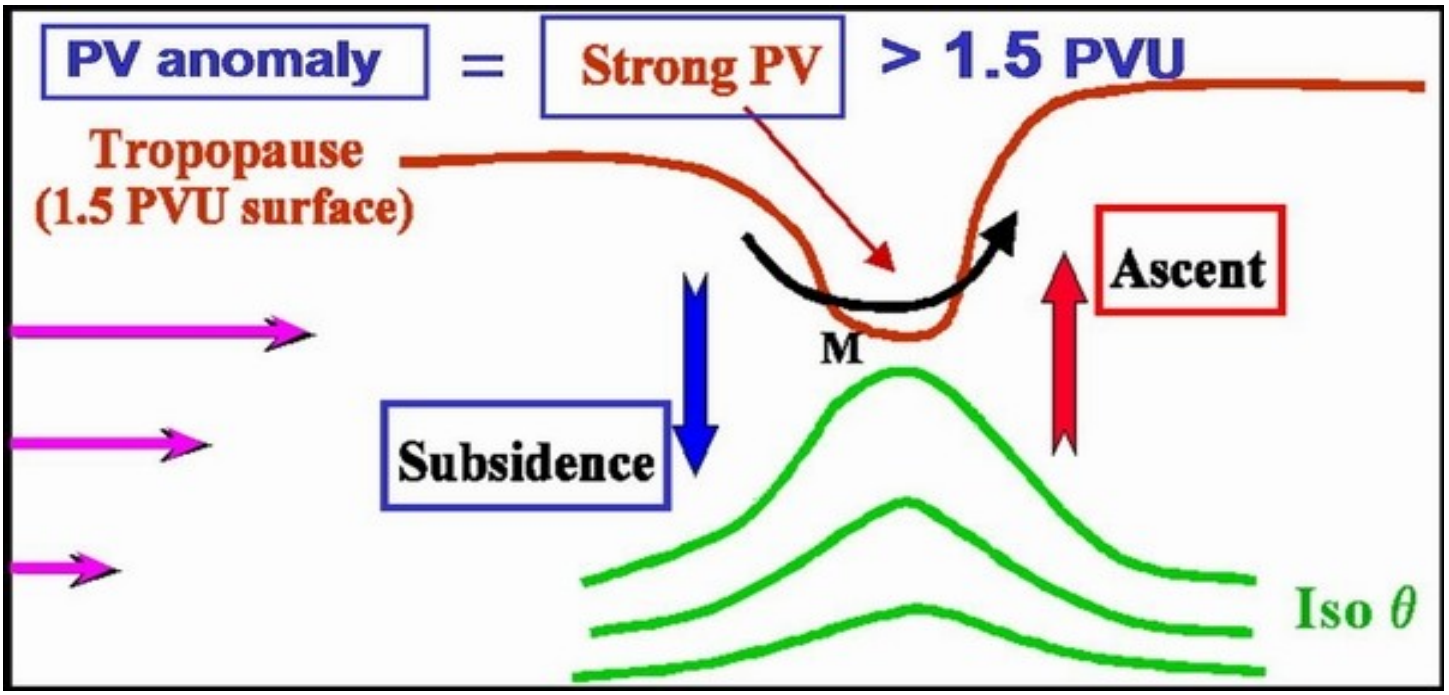

Figure 7. Dynamic tropopause anomaly (brown contour which represents the 1.5 PVU surface), and isentropes (green contours). The letter **M** denotes a geopotential height minimum. Purple vectors represent wind velocity. Note how the isentropes bend upward toward the dynamic tropopause anomaly. As the anomaly moves toward the right in this cross section, the isentropes ahead of the anomaly will rise (bend upward) toward the upper level disturbance, while those behind the PV anomaly will sink back down. The result is a pattern of ascent and subsidence associated with a traveling tropopause anomaly. Adapted from Santurette and Georgiev, **Weather Analysis and Forecasting** (Elsevier Academic Press, 2005).

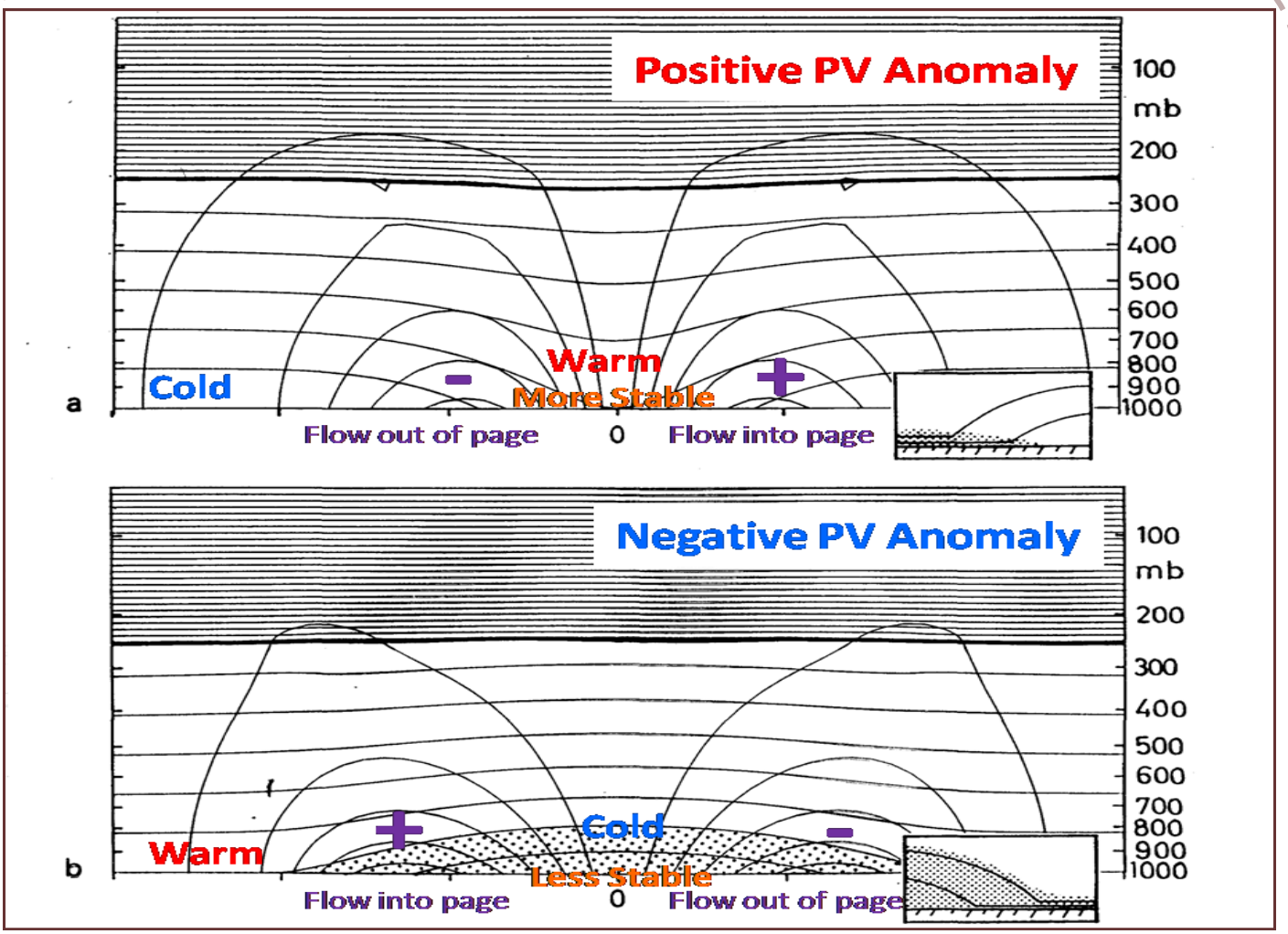

Figure 8. Idealized low level potential vorticity anomalies. Solid horizontal contours represent potential temperature, thicker black line represents the dynamic tropopause, more circular vertical contours represent isotachs (highest values near the surface). The "+" and "-" signs indicate flow into and out of the page, respectively. Top: Positive PV anomaly associated with a warm potential temperature anomaly. Bottom: Negative PV anomaly associated with a cold potential temperature anomaly. Adapted from "On the Use and Significance of Isentropic Potential Vorticity Maps", by Hoskins et al., *Quarterly Journal of the Royal Meteorological Society* (1985).

between the vertical potential temperature gradient and static stability. As isentropes bend upward ahead of an approaching dynamic tropopause anomaly, this has the effect of changing the vertical potential temperature gradient, and thus the stability of the lower layers. Specifically, this upward distortion of the isentropes results in a decrease in stability (isentropes getting pulled farther apart in the vertical). Since the stability in the low layers is decreasing, in order to conserve lower level PV, the absolute vorticity in the lower layers must increase. Thus, the upper level anomaly "induces" a low level circulation (it turns out that it also induces a circulation at higher levels as well, recall from figure 6 how isentropes above the anomaly bend downward, which decreases stability aloft). How far down into the lower troposphere this induced circulation is "felt" is dependent on stability and the scale (diameter) of the PV anomaly. An atmosphere below a tropopause anomaly that is weakly stable will allow a circulation to reach lower layers than a more strongly stable layer. In addition, given the same circulation strength, the circulation associated with a large PV anomaly will be "felt" a greater distance away than that of a smaller anomaly. This is reasonable...we don't often see surface lows spin up with weak short wave troughs, while larger waves often induce a stronger surface pressure response. Nor do we often see a strong surface pressure response with an upper wave passing over a cold arctic high and its strong low level stability. A quick way to estimate the deep layer stability of the troposphere utilizes a map of potential temperature at the 1.5 PVU surface, overlaid with low level (925/850mb) equivalent potential temperature. The smaller the difference between these two temperatures, the weaker the stability (differences less than 10 degrees K are significant).

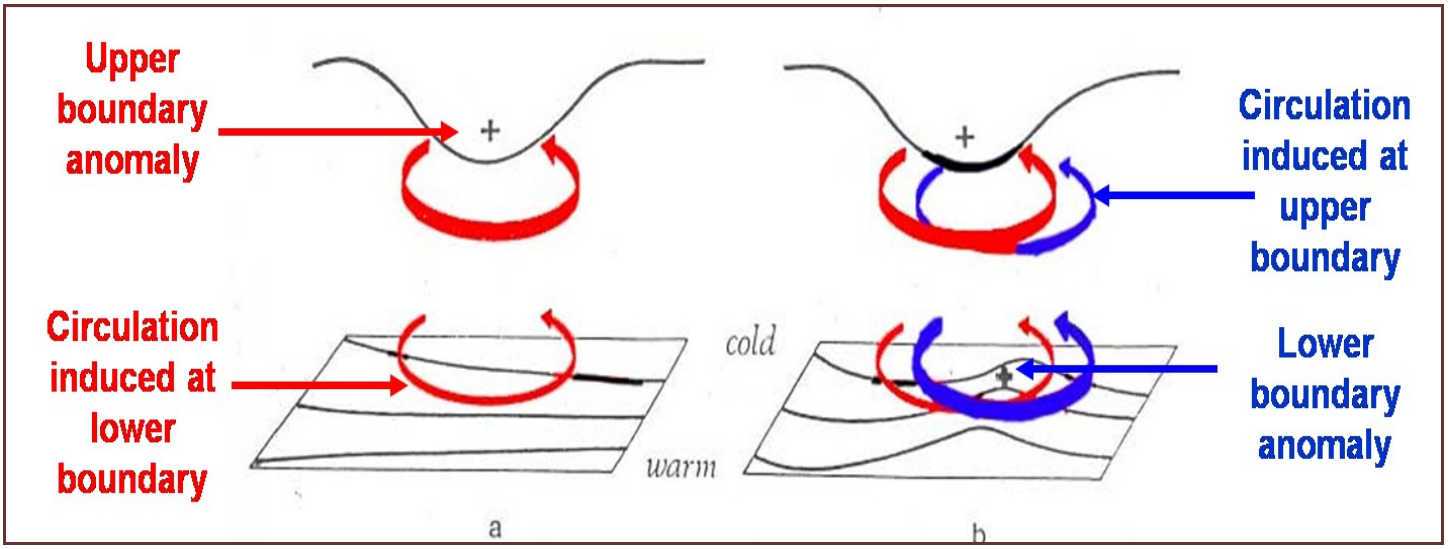

Figure 9. Cyclogenesis as a mutual strengthening of upper and lower level potential vorticity anomalies. Upper black lines represent dynamic tropopause anomaly, lower parallelogram represents a near surface level, thin contours represent isentropes with colder air to the "north" or top side of the diagram. Thick arrows represent circulations associated with the PV anomalies, thin arrows represent the circulations induced by the PV anomalies. Left: Upper level anomaly induces a low level circulation along a baroclinic zone. Right: Resultant thermal advection produces a low level thermal ridge and thus a low level PV anomaly, which then induces a circulation aloft. Adapted from Hoskins et al. (1985).

We can take this another step further and look at an upper level PV anomaly interacting with a surface baroclinic zone as shown in figure 9. On the left side of the diagram we see the circulation associated with an upper level PV anomaly and the low level circulation it induces. This low level circulation in turn alters the low level temperature gradient, given the warm air and cold air advection ahead of and behind the circulation respectively. On the right side of figure 9 we see a low level thermal ridge developing ahead of the initial circulation, which as we discussed earlier is associated with a positive PV anomaly. And in the same way the original upper level circulation impacts the low levels, the circulation associated with the low level PV anomaly then "builds upward" and strengthens the upper level circulation. So cyclogenesis can be viewed as a mutual strengthening of upper and lower level PV anomalies.

The above discussion about cyclogenesis reveals another issue regarding potential vorticity, and that has to deal with the creation of potential vorticity. In the cyclogenesis discussion, reference was made to the development of a positive PV anomaly in lower levels. But so far, all we've discussed has been with respect to the conservation properties of PV, which holds for flow that is adiabatic and frictionless. But if the flow were always adiabatic, we wouldn't have much weather since condensation, and thus cloud and precipitation production, releases latent heat, which is a <u>diabatic</u> process. Latent heat release alters the vertical stability profile, and thus can create or destroy potential vorticity, as illustrated in figure 10. A "bubble" of latent heating will increase the stability below the level of maximum latent heating, and decrease it above. These changes in stability result in an increase in PV below, and a decrease in PV above, the level of heating. This may seem confusing, since PV conservation was associated with changes in stability and subsequent changes in absolute vorticity to compensate. But here we are dealing with a diabatic, and thus non-conservative, process. So the stability changes here will result in actual changes in the value of potential vorticity. This concept shows how valuable the role of latent heating is in the cyclogenesis process. Increasing low level PV increases the strength of an existing low level circulation, as well as amplifying the circulation aloft.

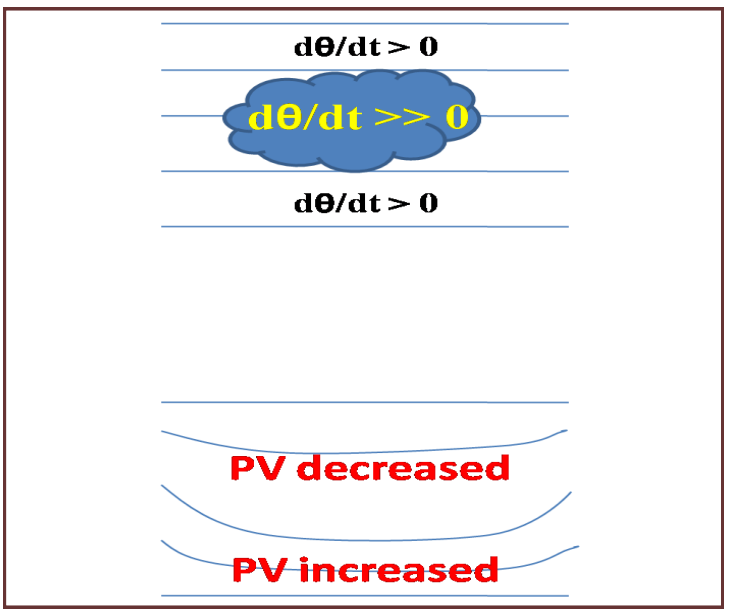

Figure 10. Top: Initial vertical potential temperature gradient (solid contours), and a "bubble" of latent heating represented by the cloud. Bottom: Vertical potential temperature gradient after latent heating, and resultant change in potential vorticity.

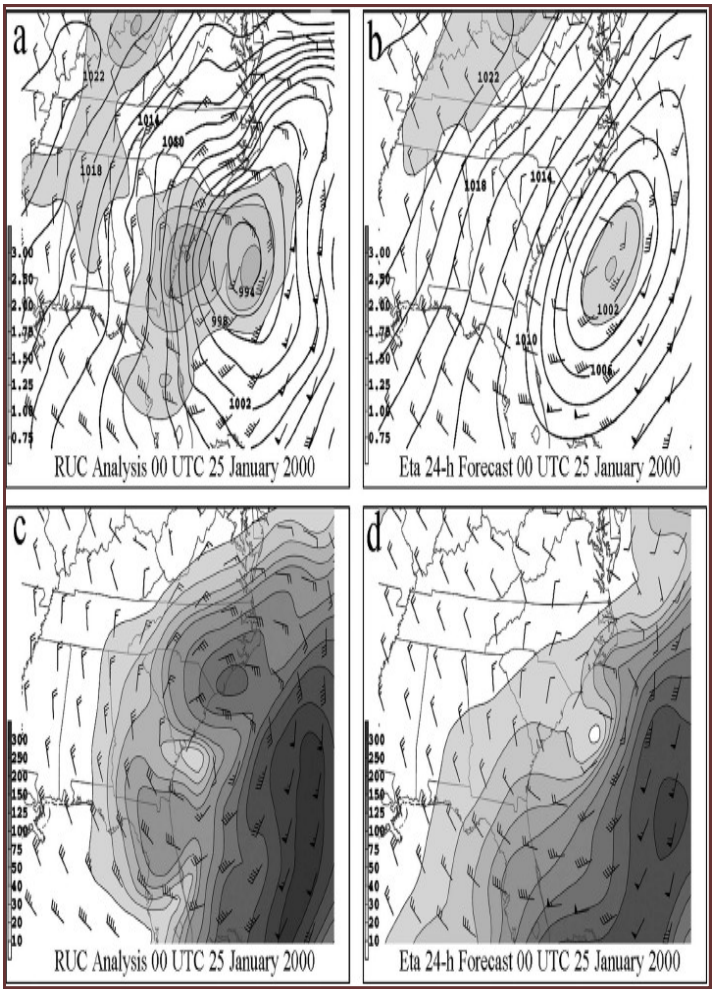

Figure 11. Top left: 00h RUC analysis valid 00z 25 January 2000 (MSLP every 2mb, 900-700mb potential vorticity (shaded), and 800mb wind barbs. Top right: Corresponding 24h Eta model forecast valid at the same time. Bottom left: 00h RUC analysis of 800mb winds and moisture flux (shaded). Bottom right: Corresponding 24h Eta model forecast valid at the same time. See text for details. From "Potential Vorticity Thinking in Operations: The Utility of Nonconservation", by Brennan et al., *Weather and Forecasting,* (2008).

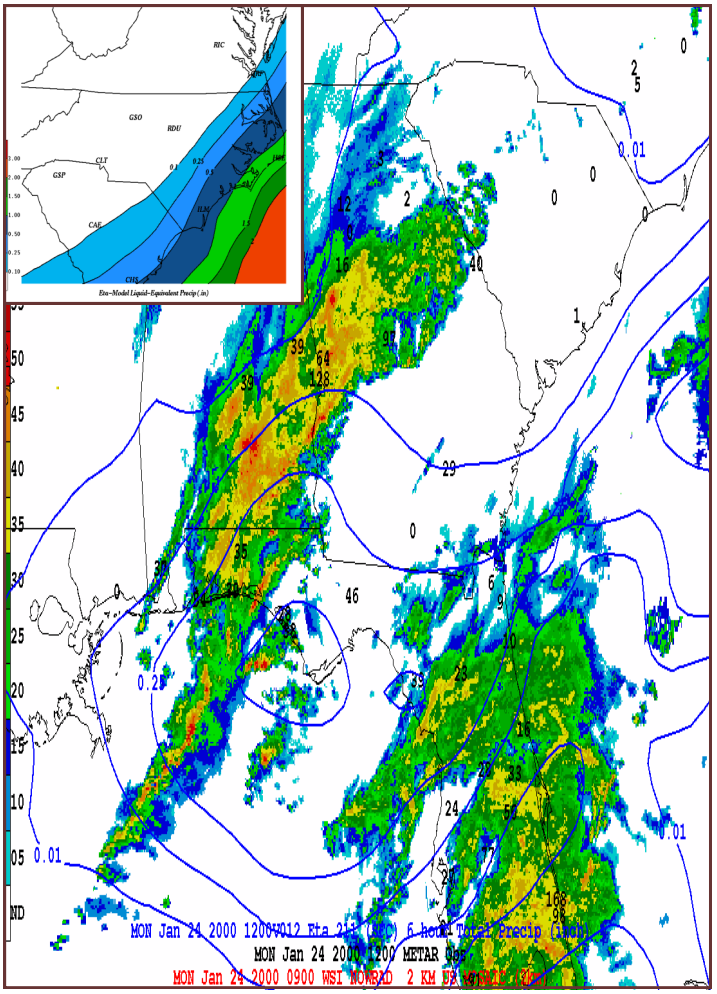

Figure 12. Eta model 6h precipitation forecast (blue contours), valid 12z 24 January 2000, radar reflectivity mosaic valid 09z 24 January 2000, and 6h observed precipitation ending 12z 24 January (black numerals). Note area of heavier rainfall across central Alabama and central and northern Georgia in an area where the Eta forecast less than 0.25 inch precipitation. Some of the observed rainfall totals in this area exceeded 1 inch. Inset: Eta model 48h total precipitation >0.10 inch ending 00z 26 January 2000. RDU is Raleigh, NC.

Another example in figures 11 and 12 above illustrates how a poor model forecast of precipitation can lead to errors in cyclone position and strength. Panels a and c show analyses of surface pressure, 900-700mb potential vorticity (shaded in panel a) and 800mb wind/moisture flux fields valid at 00z on 25 January, while panels b and d are 24h Eta model forecasts of the same variables valid at the same time. Figure 12 shows radar reflectivity, 6 hour observed precipitation, and a 6h Eta model precipitation forecast valid at 12z on 24 January. In this case, the Eta model's inability to predict an area of heavier precipitation across central portions of Alabama and Georgia, and the resultant lack of latent heating in the model, resulted in a surface low position too far offshore of the Georgia/South Carolina coast. Note the farther east and weaker 900-700mb PV field forecast by the Eta in panel b of figure 11, in comparison with the stronger and farther west PV field in the RUC analysis due to heavier than forecast precipitation that moved across Georgia and eventually South Carolina. The resultant stronger low level PV field resulted in a stronger area of low pressure developing closer to the coast than forecast by the Eta model. In addition, the low level PV anomaly enhanced the low level circulation, and resulted in a low level jet closer to the coast with a stronger feed of moisture wrapping back into the Carolinas. This led to heavy snowfall occurring across the Carolinas northward into northern Virginia that was unforecast by the Eta, including a storm total record snowfall of over 20 inches at Raleigh, NC, due in part because the model's low level moisture flux maximum did not wrap sufficiently far enough back to the west (note the Eta precipitation forecast in the inset of figure 12). Monitoring low level PV fields and comparing observed and forecast precipitation amounts allows forecasters to assess short term model behavior and the impacts of model latent heat release on forecast mass fields such as surface pressure and low level winds.

PV Application: Jet Streaks

Analyzing jet streaks on a dynamic tropopause chart gives a more complete picture of jet structure than viewing isotachs on a constant pressure analysis, since a jet streak has depth and thus will be found at more than one level. The jet stream is usually found along a gradient of potential vorticity, and jet streaks are often associated with strong dynamic tropopause anomalies, or where the dynamic tropopause is steeply sloped. Figure 13 shows a vertical cross section of potential vorticity and isotachs, showing the relationship between the dynamic tropopause and the jet core. Note the locations of the isotach maxima along the strongest potential vorticity gradient, or where the dynamic tropopause is most steeply sloped. The jet streak in the middle of the figure is associated with the polar jet, while the jet on the right side of the figure (and higher in altitude) is a branch of the subtropical jet. Figures 14 and 15 show the more defined jet structure when viewing isotachs at the level of the dynamic tropopause. Figure 14 is a map of isotachs on the 1.5 PVU surface, and shows jet cores associated with the polar and subtropical jet streams coming together across the western Atlantic Ocean. Figure 15 shows isotachs at the 250mb level, and only shows the detail of the polar jet streak, while the eastern extent and other details of the subtropical jet are not well defined...not surprising since the subtropical jet usually shows up best at the 200mb level. So viewing jet streaks on dynamic tropopause maps gives a more complete picture of the structure of jets, while viewing isotachs on constant pressure maps often only shows a slice through one or more jet streaks and may not capture the details. Coupled jet structures tend to stand out more on dynamic tropopause maps.

A couple of links where potential vorticity maps can be found online:

University of Washington

http://www.atmos.washington.edu/~hakim/tropo/info.html

University at Albany

http://www.atmos.albany.edu/index.php?d=wx_nwp

Massachusetts Institute of Technology

http://paoc.mit.edu/synoptic/forecasts/ustrop.asp

John Boris

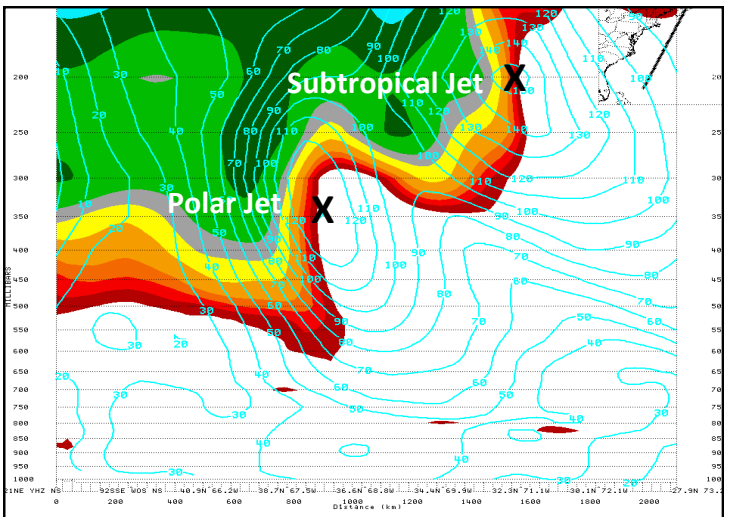

Figure 13: Vertical cross section through an upper level PV anomaly. Potential vorticity values greater than 1.5 PVU are shaded, cyan contours are isotachs every 10kt. Jet maxima are denoted by the "X"...note the location of the jet streaks where the dynamic tropopause is most steeply sloped (vertical in this case).

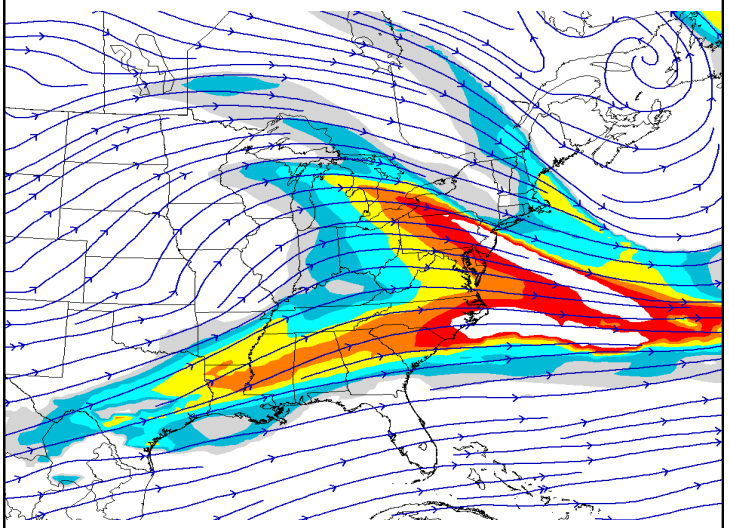

Figure 14: 1.5 PVU surface isotachs >80kt (10kt color intervals)

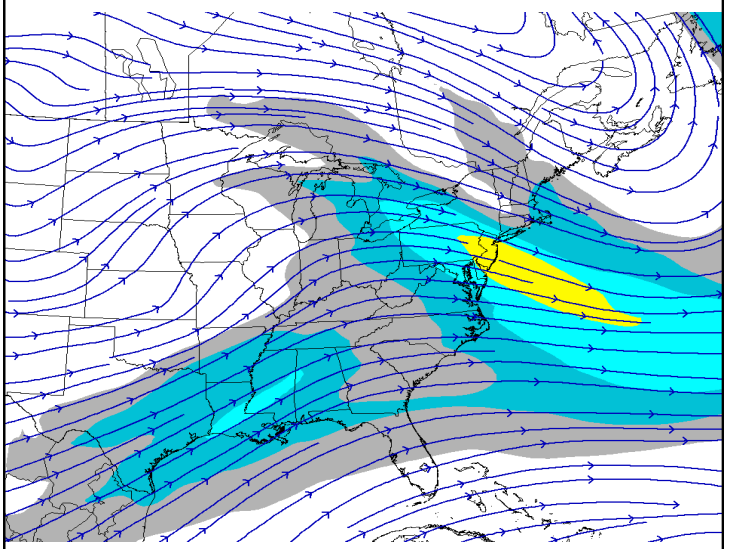

Figure 15: 250mb isotachs >80kt (20kt color intervals)