Outline

- The Dynamic Tropopause and Potential Vorticity: An Introduction

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    3. Diabatic Effects on PV
    4. Advantages of dynamic tropopause maps
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    8. Latent Heating and convective/grid-scale feedback
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The dynamic tropopause is based on potential vorticity (PV), rather than the WMO defined lapse rate definition that states (Lapse rate < 2K/km and does not exceed 2K/km for 2 km above). Isentropic surfaces making the transition from the troposphere to stratosphere usually feature a sharp gradient of Ertel PV between the two air masses (Davis and Emanuel 1991; Morgan and Nielsen-Gammon 1998). The 1-3 PVU band lies within the transition zone between the weak stratification of the upper troposphere and the relatively strong stratification of the lower stratosphere (Morgan and Nielsen-Gammon 1998). A single PV surface within this gradient is selected to define the tropopause.

The position of the tropopause is relatively flat and low at the poles, sloped upward through the mid latitudes (the typical location of the westerly polar jet in the northern hemisphere), to a relatively flat maximum height near the equator. In fact, the tropopause is so steeply sloped near upper level jets that it becomes vertical, nearly so, or even folded. The polar front jet owes its existence to a deep lower to middle tropospheric thermal gradient, whereas a typically much shallower and higher thermal gradient is found beneath the subtropical jet. Upper-tropospheric waves and the evolution of weather systems are intimately associated with excursions of the aforementioned transition region (between the troposphere and stratosphere) from its mean position (Davis and Emanuel 1991; Morgan and Nielsen-Gammon 1998). It follows that variations in the height of the tropopause are equivalent to both variations of PV near it and the amplitude of disturbance along it. The amplitude of upper-level baroclinic vortices and associated PV are usually maximized on the tropopause. Note: 1 PVU = 10^{-6} K m^2 s^{-1} kg^{-1}.

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Vertical interpolation, discussed here, is one way that dynamic tropopause maps (DT) can be constructed. Within each grid column, PV is computed upward at successive vertical levels until some threshold tropopause value – 1.5 PVU preferred here – is crossed for the last time (within the dataset domain). That level is defined as the dynamic tropopause. It is preferred to plot potential temperature and wind (barbs and isotachs) on a dynamic tropopause map, particularly since potential temperature (theta) is conserved on a constant PV surface, called an isertelic surface by Morgan and Nielsen-Gammon (1998). Such a plot reveals the advection of tropopause potential temperature and related forcing for vertical motion, as we'll discuss shortly. However, vertical interpolation allows other variables, such as pressure, to be interpolated onto that surface as well. A plot of pressure on the dynamic tropopause affords the ability to diagnose the depth of a given disturbance (ie. how far down into the troposphere the disturbance extends), independent of stability and resultant extent of vertical and lateral influence. An example of vertical interpolation at the center of an upper level trough is shown here, and reveals the dynamic tropopause (defined by the 1.5 PVU surface) extends downward into the troposphere to roughly 650 hPa. Note also that the core of the northerly upper jet that lies upstream of the upper trough axis, evident in the image at left, is centered on the 1.5 PVU surface, according to the companion cross section at right. That jet not coincidentally is also located within a sharp gradient of PV, in the vicinity of a steeply-sloped dynamic tropopause. The principle of PV conservation is useful for tracking and identification of dynamically significant features, including those driven by non-conservative diabatic processes (Brennan and Lackmann 2005; Brennan et al. 2008), which will also be discussed in subsequent slides.

Slide 5

A bundle of dynamic tropopause and related field AWIPS procedures has been created and is available for download on the WFO Raleigh Science Page (www.erh.noaa.gov/rah/science), in the Science Infusion and Sharing Section. These procedures will be introduced throughout this presentation. The list of procedures is not meant to be all-inclusive; and hopefully the material presented here will provoke some thought that will lead to the creation of additional AWIPS procedure sets in the future.

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Potential vorticity, or PV, is the product of absolute vorticity and stability. Regions of high stability (such as the stratosphere) or high absolute vorticity (such as along trough axes) are characterized by high values of PV and associated cyclonic flow. There are two important principles of PV: conservation and invertibility.

PV is conserved for adiabatic, frictionless flow; and this principle will be discussed in more detail on the following few slides.

Invertibility is the other important principle of PV. Inversion is a mathematical process of calculating all other dynamic fields from a given PV distribution. Height, wind, temperature, and even vertical motion fields can be recovered from the inversion of PV. It is possible to invert the entire PV field or an individual PV anomaly (known as piecewise inversion). PV inversion reveals the flow field induced by individual PV anomalies. In other words, it can be calculated how much of the wind at a location is caused by a given PV anomaly (given boundary conditions, a reference state, and a balance relation (Hoskins 1985)). The application of PV inversion to cyclogenesis has been used in numerous studies, including (xxx). The amplitude of the induced flow increases with size of the anomaly and with reduced static stability. The circulation induced by a PV anomaly will penetrate a certain distance above and below the anomaly, called the Rossby penetration height, given by H=f L/N (where "f " is the Coriolis parameter, " L" the horizontal scale, and "N" the Brunt-Vaisala frequency).

Morgan and Nielsen-Gammon (1998) succinctly described the importance and application of these principles to atmospheric dynamics, and reinforced the pioneer work by Hoskins 1985: "Together, conservation and invertibility imply that the dynamics of quasi-balanced flows may be expressed simply and entirely in terms of the distribution of PV and the boundary conditions necessary in the inversion of the PV. The use of conservation and invertibility to infer the dynamics of quasibalanced flows is called ''PV thinking.''" (Morgan and Nielsen-Gammon 1998).

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PV conservation requires that when a positive PV anomaly is advected from a region of high to low stability, the absolute vorticity must increase and yield consequent convergence and ascent.

Slide 8

Diabatic PV generation or destruction at the surface or tropopause will appear as a non-advective redistribution of mass. In other words, potential temperature anomalies, for example, may appear out of nowhere; and those diabatic effects can be conceptually isolated from conservative (advective) processes (Brennan and Lackmann 2005; Brennan et al. 2008; Morgan and Nielsen-Gammon 1998).

Convective feedback is manifested as spurious synoptic-scale dynamic features, including 500 hPa vorticity maxima, which suddenly appear near QPF maxima of large non-convective origin and cannot be traced back to some existing feature in the model initial conditions. Convective feedback occurs when the convective scheme fails to release enough instability to prevent the model from acting on the remaining instability on the grid scale. The model erroneously generates a significant amount of precipitation and resultant latent heat over a short period of time and over an unrealistically large area. We will investigate some of the implications of latent heat release on the surrounding environment on the following slide.

Diabatically generated distributions of PV in the lower troposphere, called internal PV anomalies, do exist; and they can alter the flow field and accordingly play an important role in moisture transport and the development of frontal waves (Brennan and Lackmann 2005; Brennan et al. 2008; Morgan and Nielsen-Gammon 1998). One of the best examples of this process was the infamous NC snowstorm of January 24-25, 2000 (Brennan and Lackmann 2005). Davis (1992) also showed that a low level PV anomaly can contribute to a significant portion of the total cyclonic circulation of a surface cyclone. Note: Surface based inversions are also considered internal PV anomalies, but are not dynamically important.

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Here is an example of a generalized, no shear case. In the presence of shear, latent heat release redistributes PV along absolute vorticity vector. In other words, shear displaces the diabatically-induced PV anomalies away from the latent heating maximum.

Height falls, convergence, and an increase in absolute vorticity, stability, and PV occur below the level of maximum heating. Conversely, height rises, divergence, and a decrease in absolute vorticity, stability, and PV occur above the level of maximum heating.

Note: Heating is not restricted to diabatic (latent) heating, and instead could pertain to a maximum of thermal advection. In other words, height (or surface pressure) falls will occur below the level of maximum warm air advection (WAA), at 850 hPa for example, according to the QG Height tendency equation. Thus, in the absence of other forcing, a surface cyclone or trough axis can often be found beneath a (baroclinic) low level jet.

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In an upper-level positive PV anomaly: the tropopause is relatively low, the upper troposphere relatively cold and unstable (indicated by the upward-sloped and relatively largely-spaced isentropes), and the lower stratosphere relatively warm and stable (indicated by the downward-sloped and relatively minimally-spaced isentropes). The static stability is therefore relatively high for an isentrope passing through the anomaly in the stratosphere, while the static stability is relatively low for an isentrope passing beneath the anomaly in the upper troposphere. Thus, upper level troughs are characterized by cold potential temperature and high pressure on the dynamic tropopause, and correspond to a low tropopause and positive (cyclonic) upper-level PV anomalies.

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In an upper-level negative PV anomaly: the tropopause is relatively high, the upper troposphere relatively warm, and the lower stratosphere relatively cold. The static stability is therefore relatively low for an isentrope passing through the anomaly in the upper troposphere, while the static stability is relatively high for an isentrope passing beneath the anomaly in the lower to middle troposphere. Thus, upper level ridges are characterized by warm potential temperature and low pressure on the dynamic tropopause, and correspond to a high tropopause and negative (anticyclonic) upper-level PV anomalies.

Slide 12

The four maps shown here all look reasonably similar and all depict flow and forcing features in the mid-upper troposphere. Each is unique, however, and has both advantages and disadvantages.

Traditional 500 hPA charts offer some forecasting advantages and will undoubtedly always have a place in the “forecaster toolbox”. Advantages include diagnoses of: *some* baroclinic shortwave troughs aloft, the magnitude of mid-tropospheric flow, mid level temperatures, and mesoscale convective vortices. Yet, 500 hPa charts have several disadvantages that make identification and tracking of baroclinic shortwave troughs difficult. These disadvantages will be highlighted on the next couple of slides.

Remember that upper-tropospheric waves and the evolution of weather systems are intimately associated with excursions of the tropopause from its mean position (Davis and Emanuel 1991; Morgan and Nielsen-Gammon 1998); and the amplitude of upper-level baroclinic waves and associated PV variations are usually maximized on the tropopause. Thus, the best way to describe and represent dynamical atmospheric processes would be with a plot of PV within the tropopause-bearing layer. This could be done by plotting vertically-integrated potential vorticity (VIP) between upper level isobaric surfaces that contain the tropopause (or with isobaric maps of upper-level PV at a single level). However, VIP maps would fail to fully capture all mid-upper tropospheric PV variations and resultantly make it difficult to identify and track all dynamic features aloft, due to the dependence upon which isobaric layer (or level) is chosen, and since PV is not advected by the horizontal wind alone. In addition, since PV is not advected by the horizontal wind alone, PV and winds on an isobaric surface or within an isobaric layer do not present a complete dynamical picture.

A better alternative would be to plot PV on isentropic surfaces which lie in the vicinity of (or cross through) the tropopause, called maps of isentropic potential vorticity. Since IPV maps display one conserved variable (PV) along the surface of another (theta), a plot of isentropic PV with wind superimposed can provide and retain the conservative advective time tendency of PV and resultant forcing for vertical motion. The path of PV -- and resultantly upper troughs -- can accordingly be traced through three-dimensional space, since PV conservation requires PV to remain confined to a particular isentropic surface (in the absence of diabatic heating). Despite the inherent dynamical information retained in isentropic potential vorticity (IPV) charts, a major disadvantage to IPV is that several isentropic surfaces are required to view the full three-dimensional distribution of PV and PV advection. The use of multiple isentropic charts to diagnose dynamic features aloft would be cumbersome and time and resource consuming - and an inefficient task in already-time-constrained operational forecasting environment.

Dynamic tropopause maps are far superior to any of the former because ***all*** PV anomalies on the tropopause are shown with ***one*** chart, regardless of which pressure level or isentropic surface they lie.

Dynamic tropopause maps are the quickest and most compact, useful, and universally-powerful way to view dynamically-significant mid to upper-tropospheric features.

Slide 13

To be fair, traditionally-used 500 hPa (H5) charts do offer some forecasting advantages, which include most notably the diagnoses of 1) deep baroclinic shortwave troughs (those which extend downward from the tropopause into the mid levels), 2) mid level height changes and associated forcing for vertical motion, 3) mid level temperatures, 4) the magnitude of mid-tropospheric flow, and 5) mesoscale convective vortices. However, these presumably most widely-used charts have major disadvantages that make identification and tracking of baroclinic shortwave troughs and associated forcing for vertical motion difficult. Use of dynamic tropopause charts for so-called, “upper level forcing”, is accordingly recommended.

The use of traditional 500 hPa charts works because most vorticity maxima at that level – those not created by diabatic heating — are associated with PV variations and equivalently disturbances at the tropopause (Morgan and Nielsen-Gammon 1998). However, there is not always a reflection of dynamically-significant tropopause disturbances at 500 hPa, namely, in the case of shallow tropopause undulations. Thus, traditional 500 hPa maps would fail to fully capture such features, if at all. An example of this is shown on the next slide.

Cyclonic vorticity maxima often disappear when passing through upper level ridges owing to the opposing anticyclonic curvature vorticity that characterized ridges, though the mid level wind maxima and shear (vorticity axes) accompanying the cyclonic vorticity maxima typically remain evident. The cyclonic vorticity then typically reappear/strengthen through an upper trough base, owing in part to the acquisition of cyclonic curvature (and to tilting frontogenesis in northwesterly flow aloft). Dynamic tropopause maps make it easier to track dynamically-significant features even through a mean ridge.

Lastly, nonconservative mass adjustments from diabatic heating – including convective feedback and gravity wave contamination in numerical weather prediction models – are usually more pronounced where latent heating is most often maximized in the mid-troposphere. The resultant diabatically-produced “noise” can make it difficult to differentiate dynamically-significant baroclinic shortwave troughs (whose maximum amplitude is attained at the tropopause) from those created diabatically. In addition, diabatically-generated mid level vorticity maxima can also get scaled up to dynamically-significant synoptic-scale disturbances in the simulated atmosphere, which can then in turn significantly alter/mask vertical motion fields through quasi-geostrophic arguments.

Slide 14

Note the tropopause undulation in the upper left panel, which extends down to about 300 hPa according to the plot of tropopause pressure, over east-central TX. The tropopause disturbance is not evident in the traditional 500 hPa analysis at top right. In fact, it is located within a 500 hPa ridge. We will investigate shortly what ramifications this feature might have on deep moist convective development when stability is low and the shallow disturbance has a larger Rossby penetration height.

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A single dynamic tropopause (DT) map can be used to identify and track through four-dimensional space dynamically important features – including upper troughs and jets – at different levels. DT maps retain the conservative advective time tendency of one conserved variable (theta) on the surface of another (potential vorticity), which not only provides dynamically correct inferences of vertical motion, but also allows the identification and evaluation of nonconservative diabatic processes such as latent heating. With an understanding of implications of diabatic processes in the real and simulated atmosphere, model diagnostics can be performed and forecast adjustments can be made in an operational forecasting environment (Brennan and Lackmann 2005; Brennan et al. 2008). DT maps can be used to diagnose and forecast forcing for vertical motion, and subsequent potential for: 1) location and intensity of cyclogenesis, 2) frontal (mesoscale) precipitation bands, and 3) moist convection, to name a few. Unlike Q-vectors, dynamic tropopause maps can be used with high resolution model data without violating QG assumptions. Thus, it is possible to view the evolution of tropopause disturbances and their associated synoptic-scale forcing for vertical motion in one to three hour increments, which also affords a better ability to assess the potential interaction between synoptic and mesoscale forcing features.

Slide 16

The depth of a tropopause undulation (or disturbance) – independent of stability and resultant extent of vertical and lateral “reach” – can be deduced from a map of DT pressure. The vertical and lateral "reach" or influence of a given tropopause disturbance is governed by the intervening stability. When a positive anomaly at upper levels passes over a region of low (high) static stability, the response and degree of induced disturbance beneath that layer is large (small). The atmosphere is more responsive to a given amount of forcing when stability is low. Lower stability promotes stronger and more focused lift (apparent in the bottom figure), and increases the probability that juxtaposed upper and lower-tropospheric forcing can interact, or couple. As a result, it is not surprising that the passage of a strong and deep tropopause undulation may often produce little sensible weather effect in winter; whereas the passage of a shallow and weak one in summer can result in significant deep moist convection, for example. Tropopause disturbances (and associated upper level PV anomalies) – whose vertical and lateral influence is governed by the intervening stability – induce flow that can modify the surrounding wind and thermal fields, even when no sensible weather impacts are apparent. So, tropopause undulations PV anomalies are not just passively advected through the flow, they instead modify that and surrounding flow.

Forecasters may get so focused on forcing mechanisms that stability often times goes overlooked. Stability should be routinely assessed in the forecast process, not just when determining the risk of thunderstorms. Saturated equivalent potential vorticity (EPV*) and lapse rates of saturated equivalent potential temperature (theta-e*) or equivalent potential temperature (theta-e) provide an excellent means of assessing atmospheric stability, particularly when the atmosphere is not characterized by static (gravitational) instability. When the atmosphere is statically unstable, more-traditional and familiar stability diagnostics such as convective available potential energy (CAPE) and temperature lapse rates (usually from H85-5 or H7-5) should be utilized. Much more research has been conducted regarding the correlation of these parameters with thunderstorm intensity.

Slide 17

This is a map of pressure on the dynamic tropopause (DT). There is an inverse relationship between tropopause height and pressure on the dynamic tropopause. High pressure on the dynamic tropopause corresponds to upper level troughs, a low tropopause, and positive (cyclonic) upper-level PV anomalies. Low pressure on the dynamic tropopause corresponds to upper level ridges, a high tropopause, and negative (anticyclonic) upper-level PV anomalies. So remember, the higher the pressure, the deeper the upper level trough.

Slide 18

This is a map of theta, or potential temperature, on the dynamic tropopause. There is a direct relationship between tropopause height and potential temperature on the dynamic tropopause. Cold potential temperature on the dynamic tropopause corresponds to upper level troughs, a low tropopause, and positive (cyclonic) upper-level PV anomalies. Warm potential temperature on the dynamic tropopause corresponds to upper level ridges, a high tropopause, and negative (anticyclonic) upper-level PV anomalies. So, upper level troughs will appear as cold areas on the dynamic tropopause, and upper level ridges will appear as warm areas. It should be noted that warm surface anomalies are also dynamically equivalent to cyclonic surface PV.

Slide 19

Differential cyclonic vorticity advection (DCVA) – or differential positive vorticity advection (DPVA) in the northern hemisphere – provide forcing for ascent; whereas differential anti-cyclonic vorticity advection (DAVA) – or differential negative vorticity advection (DNVA) in the northern hemisphere – provide forcing for descent. Remember from the previous two slides that upper level troughs are characterized by high pressure and cold potential temperature on the dynamic tropopause; and upper level ridges are characterized by low pressure and warm potential temperature on the dynamic tropopause. Thus, positive pressure and negative temperature advection are synonymous with differential cyclonic/positive vorticity advection, height falls aloft, and resultant forcing for ascent (as defined by the QG Height Tendency Equation and Isallobaric Equation). Note the strongest positive pressure (and negative temperature) advection – and contribution to forcing for ascent – over southwestern Wyoming, northern Utah, and southeastern Idaho. This is where large scale forcing for ascent and upward vertical motion should be maximized, and a likely location of precipitation provided moisture is adequate and stability is sufficiently low.

Slide 20

This is an AWIPS procedure created for comparing the dynamic tropopause between various models. The upper-level, layer-averaged absolute vorticity is double-imaged with the ECMWF (not shown for contractual reasons) and GEM-NH dynamic tropopause maps for comparative purposes, since some variables on the 1.5 PVU surface (the DT) for those models are not available below a certain level in AWIPS. Those levels are about 300 hPa for the ECMWF and 500 hPa for the GEM. As a result, no variables will be plotted in AWIPS below those levels for with those two models.

Slide 21

Steeply sloping gradients of theta/pressure on the DT correspond to a steeply sloped tropopause and resultant wind maxima. The vertical shear of the geostrophic wind is directly related to the horizontal virtual temperature gradient, based on the thermal wind relation. In other words, the maximum winds in the westerly jet stream are generally found at the tropopause, due to underlying horizontal temperature gradients.

Remember from earlier that the position of the tropopause is relatively flat and low at the poles, sloped upward through the mid latitudes, to a relatively flat maximum height near the equator. Note the sharp gradient in tropopause pressure, implied steeply-sloped tropopause, and associated jet maxima from central California to northern Colorado and southern Utah. Note also the apparent but weaker gradient in tropopause pressure and implied less steeply-sloped tropopause along the axis of the subtropical jet to the south.

Slide 22

A cross section and base line, cut from west northwest to east southeast and through an upper trough, are shown here. A positive PV anomaly and characteristic excursion of the tropopause are evident in the middle of the cross section, as indicated by the tongue or depression of locally higher values of PV. PV is contoured in black and displayed as the brighter of the two color-filled images. Also plotted is potential temperature (theta) in solid white lines, which slope upward into the positive PV anomaly and accordingly satisfy the thermal wind relation. In other words, the sloping theta contours indicate a temperature gradient to which the jet above owes its existence, based on the thermal wind relation. Isotachs into/out of the page are plotted in tan and reveal a northerly (southerly) upper jet to the west (east) of the upper trough axis (evident in the smaller of the two images above). Lastly, omega is plotted as the darker of the two color-filled, double images. The omega dipole, characterized by ascent (descent) to the east (west) of the upper trough axis (and PV max), is entirely consistent with QG omega and height tendency, and PV conservation arguments.

Slide 23

Presented here, and one of the many useful dynamic tropopause maps available in the AWIPS procedures, is a plot of tropopause pressure and wind. A single dynamic tropopause (DT) map can be used to identify upper jets located at all levels. Thus, use of several different upper-tropospheric isobaric maps – such as 200-150 hPa for the STJ and between 400 hPa and 250 hPa for the PJ, for example – is often unnecessary.

Note that the westerly jet over the northern Gulf of Mexico is maximized around 150 hPa, according to the pressure values located along the axis of the jet. In contrast, the northwesterly jet axis from central Alberta to western Nebraska and Kansas is maximized around 300 hPa to 250 hPa. Use of traditional 250 hPa or 300 hPa maps alone would not fully capture the westerly jet over the northern Gulf of Mexico, since it is centered near 150 hPa. Also note the superposition of the two jets over the west central Atlantic Ocean, indicated by the tight gradient of tropopause pressure between 150 hPa and 400 hPa. This is one case where traditional jet diagnostics with upper level isobaric maps could still be useful, since the two jet axes become superimposed and make it difficult to isolate one from the other. Such a scenario is often the case when the lower extension of the STJ overlaps and phases with the upper extension of the PJ, or when they lie parallel to one another, for example.

Slide 24

Additional charts for jet diagnostics are helpful when two jets superposition or flow within close proximity to the other, as discussed and demonstrated on the previous slide. This AWIPS procedure, created to compare dynamic tropopause wind with wind between 150 and 400 hPa, could be used in those situations. Toggled on in this particular example are 200, 250, and 300 hPa wind. Again, without the use of a dynamic tropopause chart, one would have to look at all of those various levels to locate the respective upper jets. Here we can see adjacent west southwesterly polar and subtropical jets from the Gulf Coast states to the western North Atlantic Ocean (shown well in the DT chart in the upper left panel of the 4-panel image). Note that the subtropical jet (evident on the 200 hPa plot in the lower left panel) is not depicted well by traditional 250 and 300 hPa charts (shown in the right two panels). A companion plot of dynamic tropopause pressure reveals that the polar and subtropical jets are centered near 250 and 200 hPa, respectively. The cross section shown in the image at left, whose baseline is cut from northwest to southeast and through both the polar and subtropical jets, confirms that the two jets are indeed centered at 250 hPa and at 200 hPa. These findings depict a picture entirely consistent with what we earlier learned regarding the slope of the tropopause and not coincidental vertical distribution of upper jets. Specifically, note the steep slope of the PV gradient and tropopause therein near both jets, within a more gradual upward slope from higher latitudes to lower ones.

Slide 25

The procedure shown here combines a dynamic tropopause map with maps of layer-averaged frontogenesis and stability. This procedure provides a means to diagnose the potential for coupling of upper and lower level forcing features. NAM40 data from 06Z-12Z/12 November 2009 is shown in this example. Note the positive pressure advection (top left panel), which remember is synonymous with differential cyclonic vorticity advection and associated forcing for ascent. The positive pressure advection is indicated in the same region where a sloped zone of strong and deep frontogenesis from 925 hPa to 500 hPa (right two panels) exists. It is not surprising that the axis of maximum QPF from eastern NC to central VA (plotted in the bottom left panel) is generated where upper and lower level forcing features are juxtaposed, and also where stability is weak. Stability is assessed with a layer-averaged (850-500 hPa) plot of saturated equivalent potential vorticity (EPV*) (also plotted in the bottom left panel), EPV* is a measure of all types of stability, including static, symmetric, and inertial.

Unstable conditions are present where EPV* values are negative, as indicated by dashed yellow lines. EPV* (negative and positive values) is also loaded as a double image in the same panel. For comparative purposes Q-vector convergence (850-300 hPa) can be toggled on in the upper left panel of tropopause pressure. The plot of Q-vectors demonstrates the similarities with dynamic tropopause maps with regard to inferences of forcing vertical motion. It should be noted that Q-vectors are plotted from the NAM80 model to avoid violating QG theory and concurrently creating a noisy grid. Remember, the ability to view the "dynamics" of upper level forcing features with high temporal and spatial resolution mesoscale model data is an advantage of dynamic tropopause maps relative to Q-vectors. We'll take a moment to move forward in forecast time, in three hour increments through the 6-hour period ending at 12Z.

Slide 26

For completeness, presented here is a radar loop centered around 09Z/12 Nov 2009, or right in the middle of the model forecast time period from the previous slide. Note the plume of precipitation, with an embedded mesoscale banding feature in the vicinity of the eastern NC and eastern VA state border. This precipitation axis extends across the same region of maximum 6-hour QPF indicated on the previous slide.

Slide 27

The AWIPS procedure presented here can be used to identify and evaluate nonconservative diabatic processes, such as latent heating and resultant convective feedback (in the simulated atmosphere). The identification of simulated latent heating and understanding of the effects on the surrounding environment provide a powerful model diagnostic tool. Forecast adjustments in an operational forecasting environment can then be made (Brennan and Lackmann 2005; Brennan et al. 2008).

Shown here is GFS model output for the December 18-19, 2009 middle-Atlantic snowstorm. Note the grid-scale QPF maxima (on the order of 3 to 5 inches in 6 hours) over the south central Gulf of Mexico, shown in the upper right panel. The QPF maxima and resultant latent heating consequently generates (spurious) dynamic features, including: a vorticity maxima at 500 hPa (at upper left panel), a PV anomaly at 850 hPa (at lower right panel), and a surface wave of low pressure (at lower left panel).

A positive feedback is established between the resultant spurious features and consequent QPF maxima, as each move together downstream in subsequent model forecast time steps. It should be pointed out that the development of these features in the model is consistent with what was discussed earlier regarding diabatic heating. That is, height (or pressure) falls, and an increase in absolute and potential vorticity, occur below the level of maximum heating; and height rises, and a decrease in absolute and potential vorticity occur above the level of maximum heating. Since this latent heating is occurring in the presence of vertical wind shear, the spurious 500 hPa vorticity maxima outpaces the lower level features.

It can be determined that these spurious features are an artifact of diabatic heating, since they cannot be traced back to existing features in the model initial conditions and they appear in an area void of any tropopause disturbances (according to the plot of tropopause pressure in the upper right panel). Subsequent forecast hour model data should be used with extreme caution.

The diabatic heating signals are critical in the diagnosis of the track of the incipient cyclone up the East Coast. The GFS was an outlier with respect to the track of the cyclone and indicated an offshore track; whereas the consensus of other guidance was for a solution over or just west of Cape Hatteras, NC, which will be shown on the following slides.

Slide 28

Shown here is a model comparison of 850 hPa solutions for the case presented on the previous slide. The 850 hPa PV anomaly and associated cyclone at that level are of particular importance in this case, since the total cyclonic circulation at 850 hPa was dominated and drawn eastward by the anomaly.

As shown on the previous slide, this low level PV anomaly (triggered by the intense grid scale heating) helped sustain a non-linear positive feedback cycle, whereby the resultant circulation and moisture transport led to enhanced QPF that consequently exacerbated the low level PV generation, and so on. Note in the GFS solution (at top left) the track of a spurious 850 hPa low from the central Gulf of Mexico to just southeast of Savannah, GA by 00Z/19th. The observed data did not support this spurious cyclone at 850 hPa. The cyclone in the GFS solution then tracked to a position just east of the NC Outer Banks by 12Z/19th , whereas the consensus of the other guidance – which ultimately proved correct – was for a track farther north and west, just inland of Cape Hatteras, NC.

Slide 29

Identification of the internal PV anomaly (centered near 850 hPa) was critical in the diagnosis of the track of the incipient surface cyclone up the East Coast. Remember that diabatically-generated PV in the lower troposphere (internal PV anomalies) can contribute to a significant portion of the total cyclonic circulation of a surface cyclone (Brennan and Lackmann 2005; Brennan et al. 2008; Davis 1998; Morgan and Nielsen-Gammon 1998), which was apparent in this case. In other words, PV at 850 hPa presumably "spun down" to the surface. The GFS was an outlier with respect to the track of the surface low and indicated an offshore track, whereas the consensus of other guidance was for a solution just inland of the GA coast northeastward to over or just west of Hatteras, NC. As previously noted, the GFS' offshore solution proved incorrect according to the verification analyses shown. The track of the surface low would obviously have huge ramifications on the thermal profile, QPF, and forecast snowfall across the region, to name a few. The last several minutes of discussion and accompanying slides have hopefully demonstrated the utility of diabatic heating identification, which can be assessed with the earlier introduced AWIPS procedure.

Slide 30

As previously discussed, the atmosphere is more responsive to a given amount of forcing when stability is low. It follows that shallow tropopause disturbances – even weak upper level shear axes – can trigger moist convection when the atmosphere is characterized by low static stability. According to Bosart and Lackmann (1995), the difference between tropopause potential temperature and low level equivalent potential temperature (theta-e) provide a so-called poor man's lifted index. Unlike traditional LI calculations, the poor man's version provides the advantage that (equivalent) potential temperature at both levels is conserved and unchanged by vertical motion. The advection of these variables can accordingly be inferred from the distribution of winds at each level, and future LI's can be predicted (Morgan and Nielsen-Gammon 1998).

Morgan and Nielsen-Gammon (1998), propose the use of 850 hPa theta-e, since theta is not conserved at the surface, due to diabatic heating (sensible heat flux), but it has been found that use of 850 hPa potential temperature in the “Poor man's lifted index” calculation rarely yields negative LI's and implied static instability, when used locally at WFO RAH.

Slide 31

Dynamic tropopause maps can be used to assess the qualitative magnitude and favored location of cyclogenesis. An Eady model interpretation of cyclogenesis can be obtained by combining a map of tropopause potential temperature (ϑ_T) with a map of (near) surface potential temperature, with the evolution of the tropospheric flow largely controlled by the theta distribution at each boundary (Morgan and Nielsen-Gammon 1998). “The use of charts at two levels to deduce the dynamics of the troposphere, while perhaps sounding extreme, is somewhat analogous to the traditional use of H5 and surface maps to diagnose tropospheric development” (Morgan and Nielsen-Gammon 1998).

The third term on the right-hand side (RHS) of the full (or complete) vorticity equation, the stretching/convergence (“ice skater effect”) term, mathematically describes why cyclogenesis is favored when an area of pre-existing vorticity, such as a low level frontal zone, is overspread by dynamic forcing for ascent. In addition, the second term in the thickness tendency equation – the “braking term” – is small and less restrictive to cyclogenesis when the stability is low. These ideas are vital to the positive feedback, but self-limiting, Sutcliffe/Petterssen/Forsdyke self-development cyclogenesis process, which will be reviewed on the following slide.

Slide 32

The AWIPS procedure presented here can be used to assess the qualitative magnitude and favored location of cyclogenesis, based upon the concepts and governing equations presented on the previous slide. It is a 4-panel plot that includes tropopause pressure and potential temperature (a dynamic tropopause map) with a plot of saturated equivalent potential vorticity (EPV*) to measure stability, and low level theta to measure baroclinity. Note that only two significant episodes of cyclogenesis occur during the forecast period, despite the passage of four significant tropopause disturbances through the mean eastern CONUS trough.

Cyclogenesis first takes place owing to the overspreading of a strong tropopause disturbance from the Rio Grande atop an unstable environment indicated by the negative EPV* over the western GOM, and in the vicinity of a west-east elongated low level frontal zone from the western Gulf of Mexico to the southwestern North Atlantic Ocean. Given what was just discussed regarding how these conditions affect cyclogenesis, it should be expected that the first surface cyclone should develop east northeastward along the pre-existing vorticity-rich baroclinic zone, and there should be a relatively high probability of strong cyclogenesis. Indeed, the model solution suggests such a scenario, shown in the following sequence of images. Had the model failed to depict such an evolution of cyclogenesis, it could mean we may have missed something and need to look more closely at additional data, or that the forecast solution is physically inconsistent with theory and should accordingly be questioned or discounted.

Note a trailing northern stream tropopause disturbance follows quickly on the heels of the lead southern stream wave, but additional cyclogenesis is inhibited since the axis of instability and baroclinity associated with the surface cyclone in question had already been swept eastward.

Yet another significant tropopause disturbance – shown here over North Dakota – digs southeastward but yields initially only a weak surface cyclone owing to the presence of underlying stability and a relative lack of downstream low level pre-existing vorticity (baroclinity). Rapid deepening of the surface cyclone ensues as the upper wave rounds the base of the longwave trough and overspreads a strengthening instability axis and baroclinic zone along the East Coast.

Lastly, another upper wave is apparent over the Northern Plains at the end of the forecast loop. There is no evidence of associated surface cyclogenesis, however, given the preceding stable, anticyclonic flow attendant the surface high pressure ridge across the underlying region.

Slide 33

We'll take a brief moment to review the Sutcliffe/Petterssen/Forsdyke self-development mechanism for cyclogenesis, Self-development is a self-limiting, positive feedback mechanism for cyclogenesis. The sequence of events in the self-development process begins when a pre-existing upper trough lags behind and approaches a surface front. Governed by the full vorticity equation, convergence and surface vorticity generation (spin-up) are favored along the pre-existing vorticity-rich front. The incipient surface frontal wave increases the surrounding thermal advection pattern, which serves to amplify and shorten the half-wavelength of the upper wave, according to QG height/thickness tendency arguments. In other words, warm advection ahead of the frontal wave increases thickness in the overlying ridge, while trailing cold advection decreases thickness in the trough. Diabatic heating can also significantly contribute to the amplification of the downstream ridge. A downstream ridge can be just as effective as an upstream trough in the creation of positive PV advection and resultant effects on surface cyclogenesis (Bosart and Lackmann 1995). The increased amplitude and shortened half-wavelength of the upper wave provides for increased DCVA, which in turn increases upward vertical motion and deepens the surface cyclone. The strengthened surface cyclone then promotes stronger surrounding thermal advection which in turn feeds back to the upper wave, and so on. Non-linear contributions from warm sector PBL sensible heat and moisture fluxes augment this process and can lead to explosive cyclogenesis.

Slide 34

The amplification of an upper wave, known as “compaction” of a PV anomaly or tropopause disturbance, is apparent on a dynamic tropopause map. Compaction is evident as an increase in the gradient of tropopause potential temperature and tropopause pressure along the flow direction, and as a shortening of the distance between the tropopause pressure trough and the downstream tropopause pressure ridge (Dickinson et al. 1997). Note the shortening of the half-wavelength of the upper wave with time in this example. Since PV advection is equivalent to tropopause pressure and potential temperature advection, tropopause disturbance compaction can be an important precursor of cyclogenesis (Lackmann et al. 1997). The isobaric equivalent of tropopause pressure (or potential temperature) compaction, is increasing cyclonic vorticity advection in association with increasing flow curvature (Dickinson et al. 1997).

Slide 35

The pioneer “self-development” concept of cyclogenesis described by Sutcliffe/Petterssen/Forsdyke is consistent with the more recent “PV thinking” (Hoskins 1985; Hakim et al. 1996) or mutual wave amplification perspective. That is, if a wave on an upper boundary – such as the dynamic tropopause – lags a lower boundary wave, the advection associated with each serve to amplify the disturbance on the other boundary. A positive feedback is accordingly established. Mutual amplification depends on the strength of the disturbances, their spatial scale, and the intervening stability.

Slide 36

The following are figures from Morgan and Nielsen-Gammon (1998) that demonstrate mutual wave amplification. The figures can get confusing, but once we get through them they should hopefully make some sense.

Slide 37

Presented here is another AWIPS procedure that can be used to diagnose cyclogenesis and identify mutual wave amplification. This particular example is from the cyclogenesis event from late February 2010, which brought up to several feet of snow to portions of the Northeast and Northern Mid Atlantic states. The flow associated with an upper boundary wave, characterized by cold tropopause potential temperature, can amplify a lower boundary wave characterized by warm potential temperature, provided the upper wave lags the lower one. Mimicking the figure from Morgan and Nielsen-Gammon 1998 that was presented on the previous slide, the upper two panels of this AWIPS procedure can be used to identify mutual wave amplification. The increasingly southerly component of the strong tropopause wind (plotted in the upper left panel) serves to advect the warm low level potential temperature anomaly generally northward, while the low level wind (plotted in the upper right panel) serves to advect the cold tropopause potential temperature anomaly generally eastward and around the surface low. This mutual wave amplification perspective of cyclogenesis is consistent with the concept of "self-development", as described a few slides ago. Note the waves continue to strengthen, especially apparent on the plot of MSL pressure in the lower left panel, until the two waves superimpose and the cyclone becomes vertically stacked and occludes. The occlusion process can be seen most easily in the same lower left panel, as the tropopause pressure maximum, evident by the warm colors in the image, moves atop the surface cyclone, and ultimately leads to a weakening of the surface low and surrounding baroclinity.

Slide 38

Here is a companion cross section view (cut southwest to northeast), which depicts the mutual wave amplification shown on the previous slide. Remember that the flow associated with an upper boundary wave (such as a tropopause disturbance) can amplify a lower boundary (surface) wave, provided the upper wave lags the lower one. Note the reservoir of warm potential temperature – which is equivalent to cyclonic PV – just east of the developing surface cyclone. It can be clearly seen that the cyclone is initially tilted sharply toward the southwest in the deepening, mutual amplification phase. With time, the tilt of the system weakens, the system becomes vertically stacked, and the cyclone occludes. Lastly, note how the earlier potential temperature gradient has weakened and the thermal field becomes uniform around the occluded and weakening cyclone in the last couple of images.

Slide 39

Previous slides have focused on cyclogenesis in the presence of a *single* upper level disturbance, and on *vertical* interaction with a surface wave, as described by "self-development" and the companion "PV thinking" mutual wave amplification. Observations indicate there are cyclogenesis events that occur owing to more than one upper level disturbance, including wave merger cyclogenesis (Bosart et al. 1996; Dickinson 1997; Hakim et al. 1996). *Lateral* interactions among upper level disturbances, including in the cases of 12-14 March 1993 (the Superstorm of 1993) (Bosart et al. 1996; Dickinson et al. 1997) and the 25-26 January 1978 Ohio Valley wave merger event (the Cleveland Superbomb) (Hakim et al. 1996), have also been examined. Hakim et al. found, as is shown in figure 8a, that two positive PV vortices serve to advect the other in a counterclockwise orbit about the other, with strongest interaction when they are brought into close proximity by the confluent background flow. Dark circles indicate cyclonic vortices, or PV anomalies, and open circles with positive and negative signs indicate geopotential height tendencies. They found that background flow advection dominates until the vortices are brought into close proximity. The vortices can ultimately merge and promote intense cyclogenesis, as evidence in the both the Superstorm of '93 (Bosart et al. 1996; Dickinson et al. 1997) and the Cleveland Superbomb (Hakim et al. 1996). Morgan and Nielsen-Gammon (1998) examined a case (4-6 November 1988) where instead of merger, a lead trough lifts out while an upstream one amplifies southward and replaces the former. In that case, the eastern trough ''induced'' northerlies within, and accordingly amplified, the western trough; whereas the western trough induced southerlies within, and accordingly weakened, the eastern trough.

Slide 40

Presented here is an example of lateral PV interaction during the intense Veterans Day (11-13 November 2009) Nor 'Easter. Initial numerical weather prediction solutions (not shown) indicated a southward suppressed surface cyclone, owing to the swift eastward movement of the disturbance in the southern stream relative to the one in the northern stream, and resultant separation and lack of interaction between the two. However, the trend in the NWP models was for apparent increased lateral interaction, eventual trough merger, and consequent much slower and farther northward track of the attendant surface cyclone. It is proposed that the PV anomaly in the southern stream was stronger and resultantly induced stronger northerlies within, and accordingly amplified, the one in the northern stream. The resultant more amplified northern stream PV anomaly was then able to “capture” the one in the southern stream as the two were brought into closer proximity. This hypothesis could be tested through PV inversion. The amalgamation of the two vortices of interest and subsequent merged synoptic scale trough underwent a change from positive (meridional) tilt prior to cyclogenesis to negative tilt during intense cyclogenesis, which is entirely consistent with QG energetics (not described in this lesson).

Slide 41

Forecasters must be cognizant of and understand how to identify the following dynamic tropopause map caveats. Note that there is a void of tropopause data from the Great Lakes to the lower Mississippi valley on the dynamic tropopause map. Remember, PV is computed upward at successive vertical levels until the threshold tropopause value – defined here as 1.5 PVU – is crossed for the last time within each grid column. If PV is less than 1.5 PVU throughout the dataset domain of a particular grid volume, as is apparently the case in this example, then the 1.5 PVU isertelic surface and variables plotted thereon do not exist in the model. This idea can be confirmed with the companion cross section at right, whose baseline was cut from west to east through the full latitude trough over central North America and a downstream ridge over the East Coast. The cross section reveals the 1.5 PVU surface indeed curves upward and out of the dataset domain.

If an internal PV anomaly (>1.5 PVU in this case) exists in the data, and if the dynamic tropopause is high and exists above the upper limit to the model dataset domain, then the pressure level of the internal PV anomaly would be plotted. It is important to identify where this occurs, since the internal PV anomaly could otherwise be interpreted as a significant tropopause disturbance with downward extent well into the lower troposphere.

Slide 42

Rather than having to listen to me any longer, please read through these summary slides on your own. While not all-inclusive, they do capture most of the main points of what has been covered.

Summary

- DT maps can be used to diagnose/forecast forcing for vertical motion and subsequent potential for cyclogenesis, frontal precipitation bands, and moist convection.

- Upper level troughs (ridges) are characterized by cold (warm) potential temperature (θ)and high (low) pressure on the DT, and correspond to a low (high) tropopause and positive (negative) upper-level PV anomalies.
- Positive (negative) $pressure_{Tropopause}$ advection and negative (positive) $\theta_{Tropopause}$ advection are synonymous with DCVA/height falls aloft and resultant forcing for ascent.
- Steeply sloping gradients of θ/pressure on the DT correspond to a steeply- sloped tropopause and resultant wind maxima.

Summary

- Atmosphere more responsive to a given amount of forcing when stability is low. Thus, shallow tropopause disturbances – even weak upper level shear axes – can trigger deep moist convection when the atmosphere is unstable.
- Other advantages of DT maps are the interpretation of cyclogenesis by: Eady model, Sutcliffe/Petterssen/Forsdyke self-development, and "PV thinking".
- Vertical and lateral PV interactions are possible, and governed by strength of the disturbances, their spatial scale, and the intervening stability.
- DT map failures include when the 1.5 PVU surface is located above the highest model vertical level and when an internal PV anomaly exists in the same instance.

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Questions

Links to this presentation, the associated presenter notes, and the AWIPS procedures are available from the NWS Raleigh Science Program page.