1. IC5.1: Diagnosing Synoptic Scale Internal Forcing: A Review of QG Theory and Potential Vorticity

Instructor Notes: Welcome to the AWOC Winter Track Instructional Component 5, Lesson 1. Phillip Schumacher and Pat Market are the presenters for this presentation, entitled Diagnosing Synoptic Scale Internal Forcing: A Review of QG Theory and Potential Vorticity, which should last approximately 35 to 40 minutes.

Student Notes:

2. Lesson Objectives

Instructor Notes: By the end of this lesson, you should understand the following topics of quasi-geostrophic theory and Q-vectors: appropriate grid resolution, the relationship to thermal gradients, and advantages for using in operations. You should also be able to address the following issues with potential vorticity and tropopause maps: identifying synoptic scale features, anticipating movement and development of waves, determining areas of cyclogenesis and vertical motion, and anticipating impacts on low-level wind fields.

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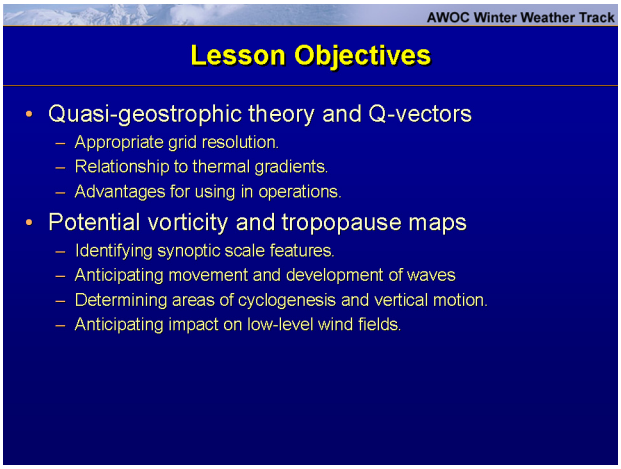
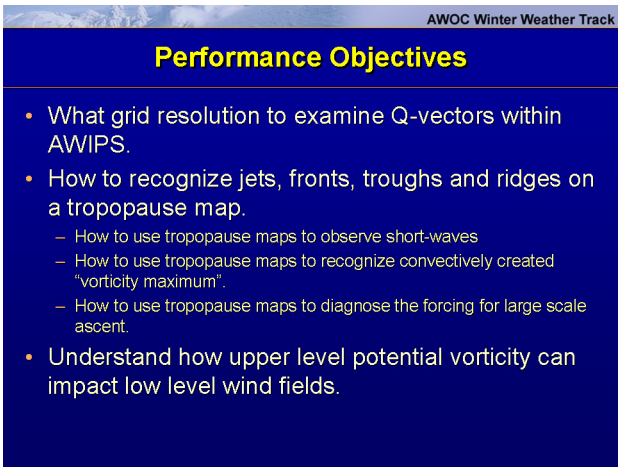


3. Performance Objectives

Instructor Notes: There are several performance objectives for this lesson. After completing this lesson, you should be able to demonstrate the following: the proper grid resolution to examine Q-vectors within AWIPS; how to recognize jets, fronts, troughs, and ridges on a tropopause map (including how to use tropopause maps to observe short-waves, recognize convectively created “vorticity maximum”, and diagnosing the forcing for large scale ascent); and understanding how upper level potential vorticity can impact low-level wind fields.

Student Notes:



4. What Are Two Assumptions Made in Deriving the Quasi-Geostrophic Equations?

Instructor Notes: We realize that many of you have seen QG-theory somewhere before. Therefore, we will not spend a lot of time on the details of QG-theory and how to apply it operationally. Rather, this lesson will discuss how to best apply QG-theory to mesoscale forecasts of winter weather. To achieve this, we will ask you a series of questions. If you get the question correct, you will be given the option to hear an explanation of the correct answer. If you answer incorrectly, you will automatically be taken to an explanation of the correct answer. Here is your first question. What are two assumptions made in deriving the quasi-geostrophic equations?

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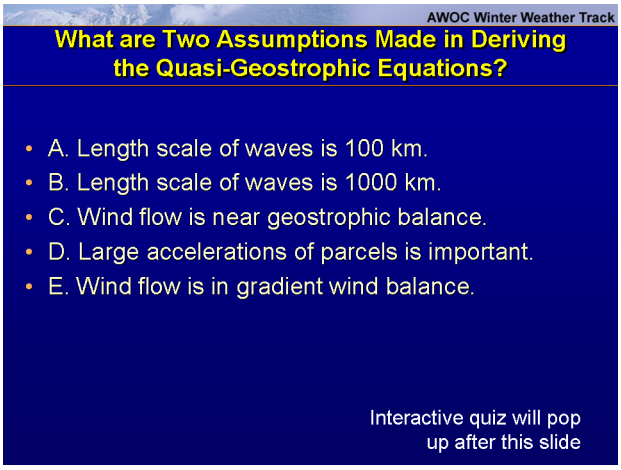


5. Assumptions in QG Forcing

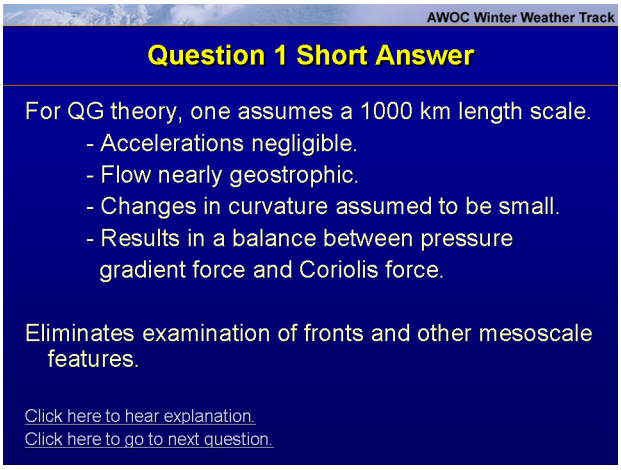
Instructor Notes: Interactive quiz question

Student Notes:

6. Answer to Quiz Question #1

Instructor Notes: When the quasi-geostrophic equations are developed, several assumptions need to be made. One assumption is that the horizontal length scale of any disturbance is around 1000 km. Two other assumptions are then made: that the mean geostrophic wind speed is 10 m/s and the mean ageostrophic wind speed is 1 m/s. When this is done, then one can scale the primitive equations to remove parcel acceleration and the ageostrophic wind. Then through a lot of mathematical manipulation we can get the quasi-geostrophic omega and height tendency equations. However, by using these assumptions we have eliminated an examination of all mesoscale features such as fronts as well as highly curved flows where the role of ageostrophic flow is large. Therefore QG theory and the QG equations only apply for synoptic scale features.

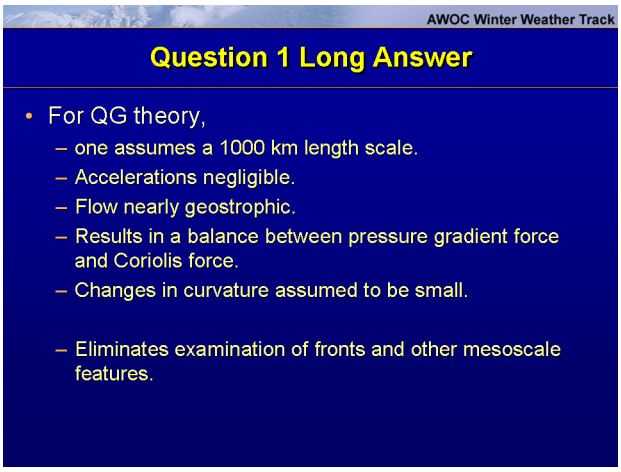
Student Notes:



7. Long Answer to Quiz Question #1

Instructor Notes: The instructor provides a more detailed explanation for the answer to the quiz question in the presentation.

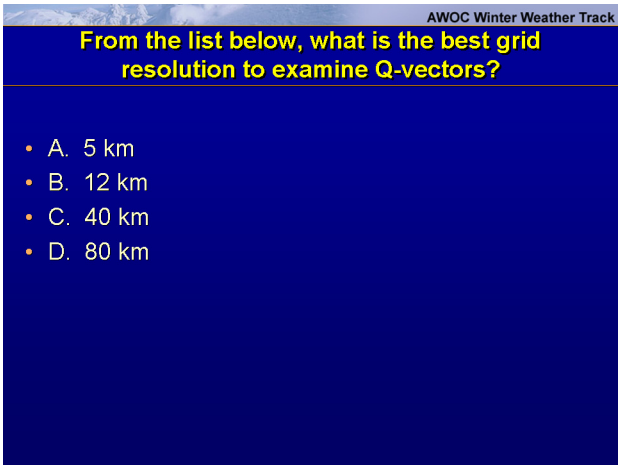
Student Notes:



8. From the List Below, What is the Best Grid Resolution to Examine Q-Vectors?

Instructor Notes: Here's your second question. From the list below, what is the best grid resolution to examine Q-vectors?

Student Notes:



9. Best Grid Resolution for QG

Instructor Notes: Interactive quiz question.

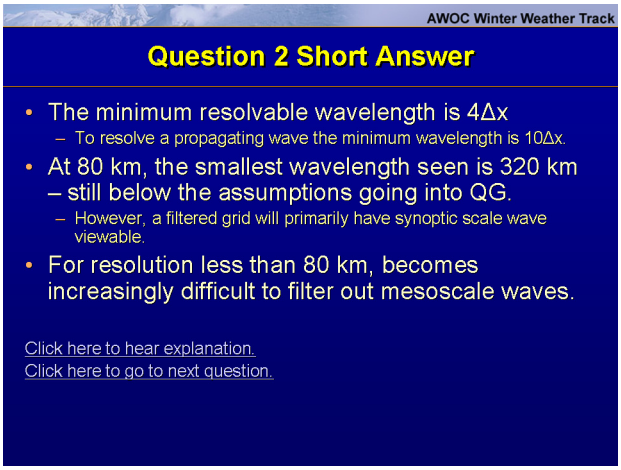
Student Notes:

10. Answer to Quiz Question #2

Instructor Notes: The correct answer is D. A standing wave can be resolved to 4 times the grid spacing. However, to capture a propagating wave, the smallest resolvable wave would need to be 10 times the grids spacing. So, looking at the answers provided, in a 5 km grid the smallest resolvable wave would be between 20 and 50 km, for a 10 km grid between 40 and 100 km, for a 40 km grid between 160 and 400 km and for an 80 km grid between 320 km and 800 km. Both the 5 and 10 km grid resolve wavelengths well below the assumptions used to derive the QG omega equation. Even a 40 km grid will resolve waves which are more mesoscale than synoptic scale. Since we only want to use Q-vectors on a grid scale which is consistent with the assumption made to derive the equation, then one needs to have a 80 km grid. Other mesoscale grids will include energy from waves on the mesoscale and therefore not provide information on the synoptic-scale forcing for ascent. In fact, even an 80 km grid is on the low end for calculating Q-vectors.

Barnes et al. (1996) showed that noise in the height field of these models can feed back into calculations of the geostrophic wind and QG theory. Barnes developed a filter to use on these type of grids which can remove much of the mesoscale energy but retain the synoptic-scale wave energy. This filter is generally applied to an 80 km grid to remove these small scale waves. This filter cannot be applied successfully on smaller grids such as 40 km, however. The filter would need to be applied several times to remove mesoscale energy. Barnes has shown that when applied numerous times, it will also remove a significant amount of energy for synoptic-scale waves.

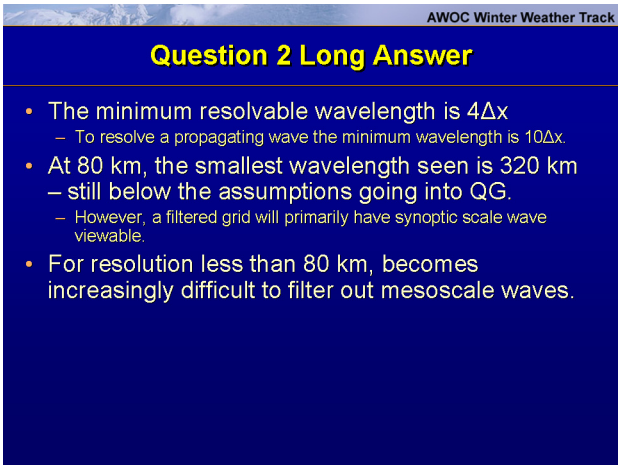
Student Notes:



11. Long Answer to Quiz Question #2

Instructor Notes: The instructor provides a more detailed explanation for the answer to the quiz question in the presentation.

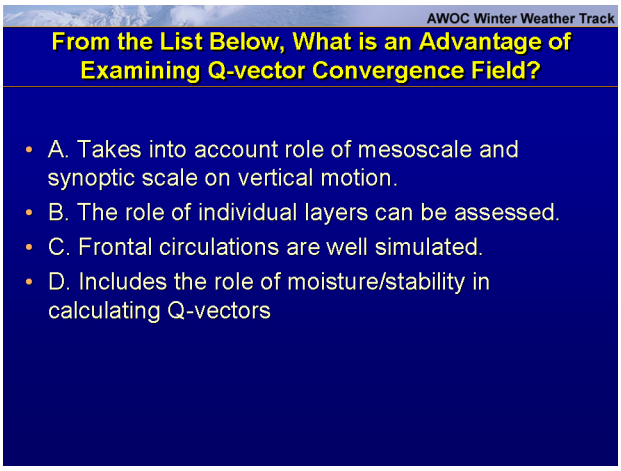
Student Notes:



12. From the List Below, What is an Advantage of Examining Q-Vector Convergence Field?

Instructor Notes: Question #3. From the list below, what is an advantage of examining the Q-vector convergence field?

Student Notes:



13. Advantage of Q-Vector Convergence

Instructor Notes: Interactive quiz question.

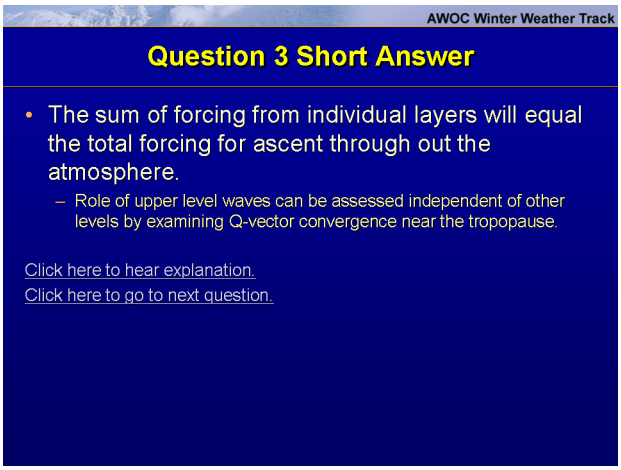
Student Notes:

14. Answer to Quiz Question #3

Instructor Notes: The correct answer is B. The sum of the forcing at individual layers will equal the total forcing through the entire troposphere. A is not correct because Q-vectors can only be used to assess synoptic-scale forcing. The assumptions used to derive the equations do not allow for examination of mesoscale waves. C is not correct because the cross-front circulation is a mesoscale phenomenon. While not shown in this

talk, the QG frontal circulation is not realistic. D is not correct because Q-vectors themselves do not include spatial variations, stability, and moisture. However, were one to solve the QG omega equation, stability does play a role. When you examine Q-vectors and its convergence you are only examining the ageostrophic response to geostrophic advection. B is correct. You can examine an individual layer and determine that layer's influence on the overall QG-vertical motion. Therefore, one can look at a layer near the tropopause and determine its influence on the overall synoptic scale forcing for vertical motion. This does not say there will be ascent or subsidence. Rather it tells you that were this layer acting alone there would be ascent through the troposphere. As we will discuss later in this lesson, this can also be done with potential vorticity – however, the local WFO lacks the computer power or time to do this type of calculation. In Lesson 2, we will discuss how this property of Q-vectors can be used to determine coupling between the upper-level wave and low- or midlevel frontal circulation.

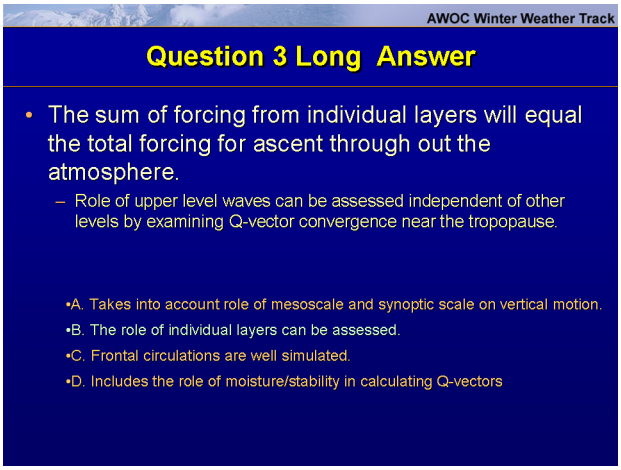
Student Notes:



15. Long Answer to Quiz Question #3

Instructor Notes: The instructor provides a more detailed explanation for the answer to the quiz question in the presentation.

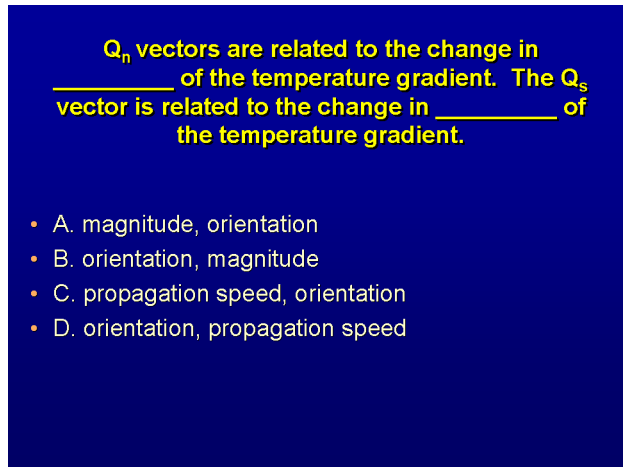
Student Notes:



16. Q_n and Q_s Vectors Question

Instructor Notes: And finally, here's your last question. Q_n vectors are related to the change in "blank" of the temperature gradient. The Q_s vector is related to the change in "blank" of the temperature gradient.

Student Notes:



17. Q-Vectors & Temperature Gradients

Instructor Notes: Interactive quiz question.

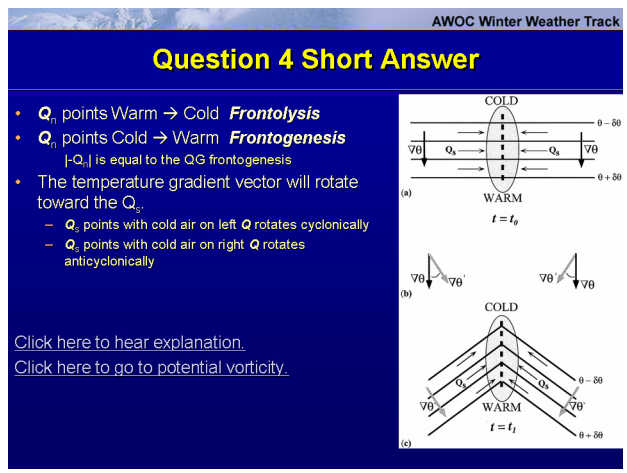
Student Notes:

18. Answer to Quiz Question 4

Instructor Notes: A is the correct answer. Q_n and Q_s do not tell you anything about the movement of a frontal boundary. However, they do provide information on whether a thermal gradient is frontogenetic or frontolytic and which way the isotherms are rotating. In most cases, forecasters are concerned with Q_n and whether a thermal gradient is frontogenetic. When the Q-vector points from warm air to cold air then the gradient is fron-

tolytic. When pointing from cold air to warm air, then it is frontogenetic. In addition, the magnitude of the Q_n vector is equal to the geostrophic frontogenesis. We'll discuss geostrophic frontogenesis in more detail in Lesson 2. The Q_s vector is showing changes in orientation to the thermal gradient. While used less often in operational forecasting, the diagram from Martin (1999) shows that the Q_s forcing plays a role in reorienting isotherms to produce a trowal structure within the occlusion of a cyclone. This will be discussed in lesson 5. What is also seen is that, when Q_s is convergent along a thermal gradient, it will act to rotate the isotherms in such way that a thermal ridge will develop on a front where the convergence is maximized. When Q_s is divergent along a thermal gradient, the geostrophic wind is acting to rotate the isotherms so that a thermal trough will develop on the front. The thermal trough will develop where the divergence is maximized along the gradient.

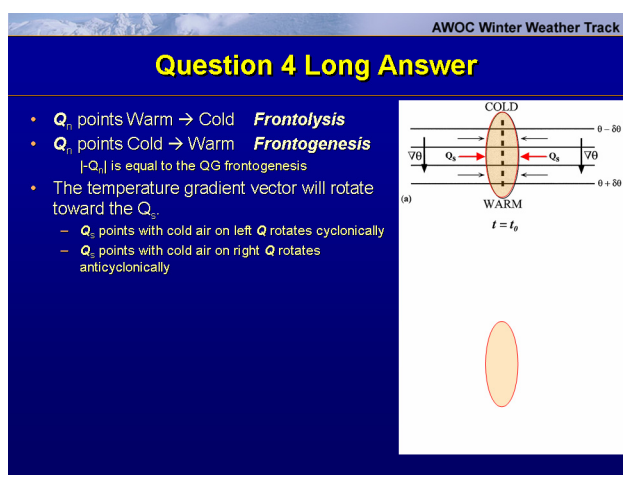
Student Notes:



19. Long Answer to Quiz Question 4

Instructor Notes: The instructor provides a more detailed explanation for the answer to the quiz question in the presentation.

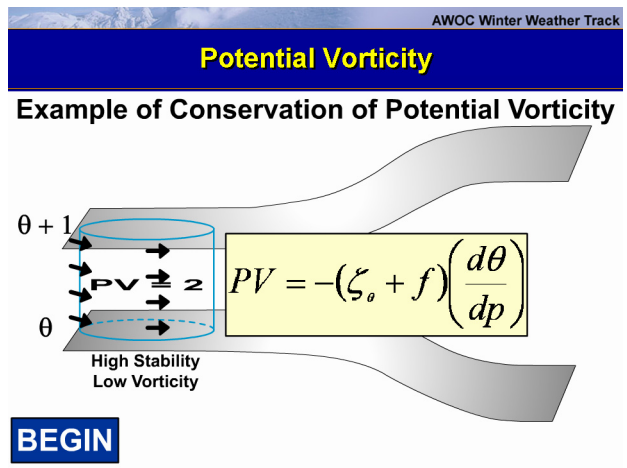
Student Notes:



20. Potential Vorticity

Instructor Notes: Potential vorticity (PV) is the product of the absolute vorticity and the static stability. Typically potential vorticity is calculated on an isentropic surface since PV is conserved in motions along an isentropic surface. PV will be positive in an inertially stable and dry statically stable atmosphere. Except for brief periods of time over small areas, this is true. Because PV is conserved in motions along isentropic surfaces we can anticipate changes in the vorticity if we examine the advection of potential vorticity in a fluid. In the animation above, we start with a parcel in a stable environment. As the parcel moves to a less stable environment, there is no change in PV which means that the vorticity must increase. By the time, it reaches a weakly stable environment, the vorticity has increased markedly and one can imagine the development of a strong cyclonic circulation.

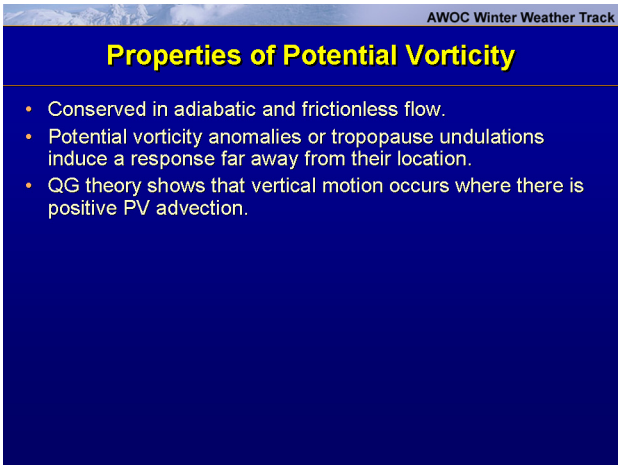
Student Notes:



21. Properties of Potential Vorticity

Instructor Notes: Potential vorticity is of interest to forecasters for two reasons. First, potential vorticity is a conserved quantity in adiabatic and frictionless flow. This makes it easier to follow short-waves, which have high values of PV, as they move through troughs and ridges. One can also picture how wave will be advected and distorted in a particular wind field. Second, potential vorticity anomalies (i.e. troughs or ridges) induce a response in the wind field away from the anomaly. While this fact has been known for a while, one can mathematically show how these anomalies affect the wind and temperature field. This relationship has made it easier to develop conceptual models which allow forecasters to anticipate how the wind field may adjust to an approaching wave. We can picture how an approaching wave may affect the low-level wind field and also where the lift or cyclogenesis is most likely to occur. Later in this lesson as well as in Lesson 2, we will use PV inversion to help understand how upper-level waves can impact low level frontal circulations.

Student Notes:



22. Tropopause Maps

Instructor Notes: The biggest benefit to potential vorticity in operations is the ability to look at the dynamic tropopause on a tropopause map. Above is a comparison of a tropopause map with potential temperature and wind and the 300 mb surface with isotachs and height. The concept of a tropopause map is similar to that of a isentropic map. However, instead of plotting variables onto a constant potential temperature surface, we plot variables (wind, temperature, etc.) onto a constant potential vorticity surface. While there are occasions where potential vorticity can have the same value at different heights, in practice, the lowest pressure where that value occurs is usually plotted. These plots are called tropopause maps because the 1.5 or 2.0 PVU surface is used which is typically defined as the dynamic tropopause. Above 2.0 PVU it is assumed the air is of stratospheric origin. So why use tropopause maps? The reason is that many features can be observed quickly. Like PV on an isentropic surface, potential temperature on a PV surface is conserved. While pressure is not conserved, it can also be plotted on tropopause maps and there are advantages to doing so. The biggest advantage is being able to see how low into the atmosphere a particular short-wave, or PV anomaly, descends into the troposphere. As we will discuss in Lesson 3, this observation has an influence on how well an upper wave can interact with a low-level front. Further, since the tropopause varies with height, one can usually see both the polar and subtropical jets on the same map. This relationship is generally not true on either isentropic or constant pressure maps. Other features of tropopause maps include: high pressure (cold potential temperature) which indicates positive PV anomalies, tight pressure (thermal) gradients are indicative of jet streaks, positive pressure (negative potential temperature) advection can indicate wave/jet movement, and positive pressure (negative potential temperature) advection is associated with large scale forcing for lift. Finally, because potential temperature is conserved on PV surfaces in the absence of convection, it is possible to follow individual PV anomalies and to anticipate their future movement based upon the wind field on the tropopause.

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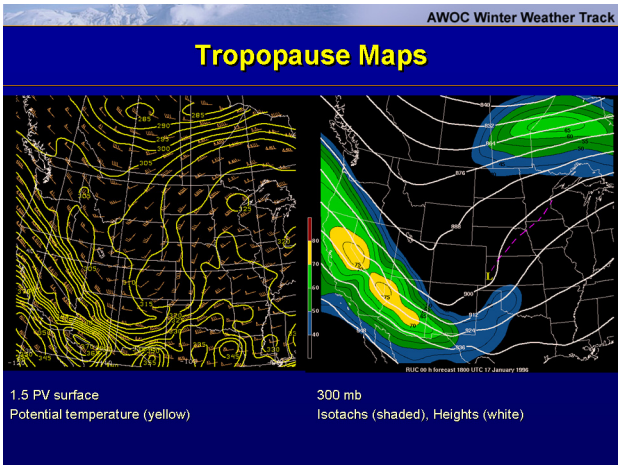
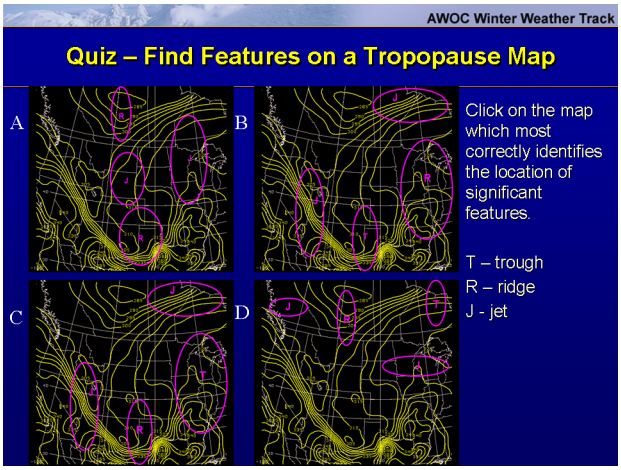


23. Quiz – Find Features on a Tropopause Map

Instructor Notes: Let's test your interpretation of tropopause maps. There are 4 tropopause maps shown with several ellipses. Each ellipse has a letter in it indicating it is circling a jet (J), ridge (R), or trough (T). Click on the image in which the jet, trough and ridge are labeled correctly.

Student Notes:



24. Potential Temperature Question

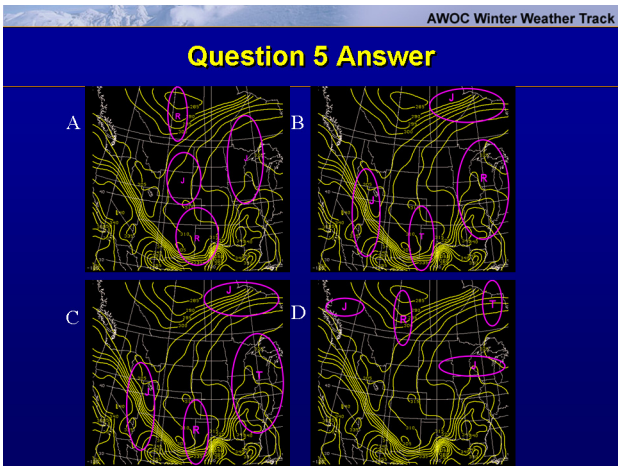
Instructor Notes: Interactive quiz question.

Student Notes:

25. Answer to Quiz Question #5

Instructor Notes: The jets are located where the potential temperature gradients are largest: one over the western United States from Oregon to New Mexico and the second over southern Canada from Manitoba to Hudson Bay. The trough is located where there is a relative minimum in potential temperature. The most pronounced trough is over western Texas, although there is a second trough located along the Alberta and Saskatchewan border. Finally, the ridge is located where there is a relative maximum in potential temperature (along the Mississippi River Valley). Note that there is also likely a ridge off the west coast of the United States.

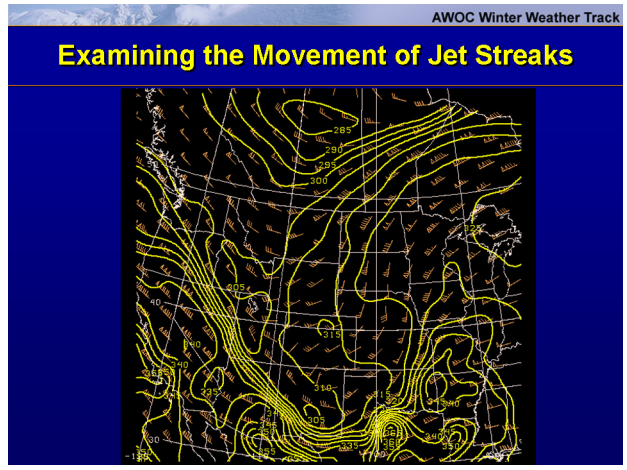
Student Notes:



26. Examining the Movement of Jet Streaks

Instructor Notes: With tropopause maps, we can also examine if jets will propagate or remain nearly stationary. Let's examine the jet streak across southern Canada which is indicated by the tight potential temperature gradient extending from southwest Saskatchewan into Ontario. Notice that northwest of the thermal gradient there is cold advection while to the south of the gradient there is neutral or even weak warm advection. Since this is winter, we will assume that convection is not significant and potential temperature is conserved. Since potential temperature is conserved this would imply that the thermal gradient should intensify across Manitoba which would hold the entrance region of the jet in place.

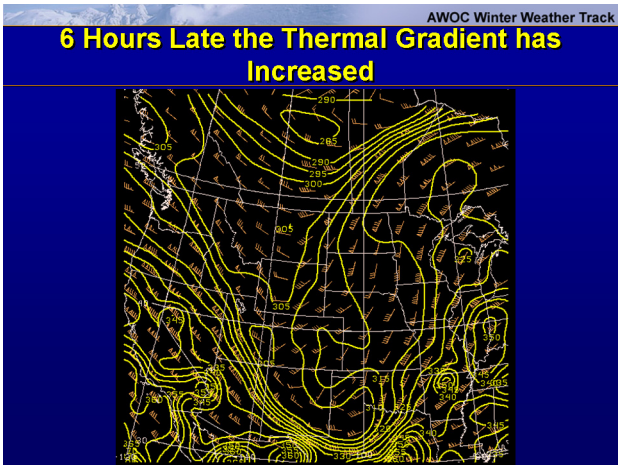
Student Notes:



27. 6 Hours Later the Thermal Gradient has Increased

Instructor Notes: Six hours later we can see that, in fact, the thermal gradient on the tropopause has tightened and remained nearly stationary across southern Manitoba and the jet streak has progressed very little downstream over the six hours. In fact, the entire long wave system has amplified as the southern storm trough has moved into western Oklahoma. A couple of observations examining jet streaks on tropopause maps: Recall that we can imply large-scale forcing for ascent when there is cold (positive pressure) advection on the tropopause. As discussed earlier, there are two jet streaks on the map above. We have already discussed the northern jet streak. Notice that, when using a contour interval of 5 K, that the cold advection is not readily apparent within the entrance region over the Dakotas and Minnesota – especially 00Z 18 January (previous slide). However, with the southern jet streak, extending from Utah into Texas, we can see a tight gradient in the left exit region of the jet streak. Both areas are a focus for lift. In fact, there was a significant blizzard and ice storm over the northern plains and Great Lakes. Yet, the advection within the entrance region of the northern jet appears less significant. Displaying contours of wind speed with wind and pressure (or potential temperature) on the tropopause can help one recognize areas where advections on the tropopause may be more subtle, especially with jet streaks. Also, as we will discuss in Lesson 2 and 3, fronts and stability can play a large role in determining the impact of a PV anomaly on vertical motion. Typically, the anticyclonic shear side (south side) of a jet streak is an area with lower stability.

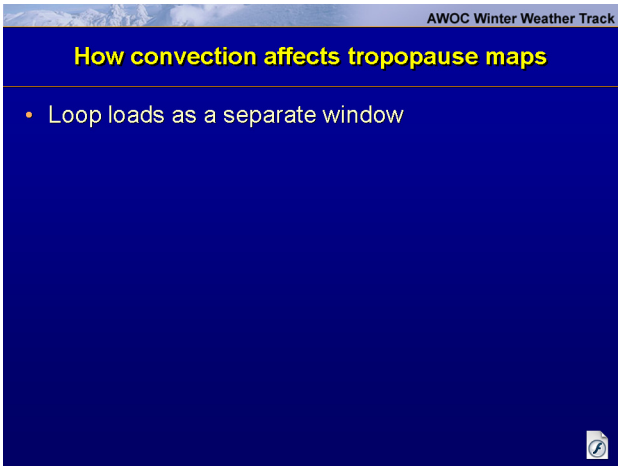
Student Notes:



28. How Convection Affects Tropopause Maps

Instructor Notes: While this course in not about convection, tropopause maps can also be used to determine whether a vorticity maximum seen at 500 mb is convectively produced or part of a baroclinic wave. We will examine two loops. The first will show 3-hour QPF and 500 mb vorticity for the first 15 hours of the Eta forecast beginning at 0000 UTC 24 June 2003. As you look at this loop, notice that the vorticity rapidly increases just ahead of where the QPF is maximized. After you finish examining that loop, click on the toggle and examine the same loop with tropopause pressure and wind with 3-hour QPF. Notice that where the 500 mb vorticity maximum was, the tropopause pressure decreases rapidly. This is not a result of advection as there is no lower pressure on the tropopause upstream. Instead, the pressures are decreasing where the QPF is maximized. This is a result of non-conservation. The latent heat release associated with convection “destroys” potential vorticity. This results in raising the tropopause locally. Then this area of lower pressure is advected downstream resulting in the development of a mesoscale ridge aloft. Notice the development of this ridge is where the 500 mb vorticity is maximized. By overlaying tropopause pressure with 500 mb vorticity, one can determine whether a particular vorticity maximum is baroclinic or convectively induced.

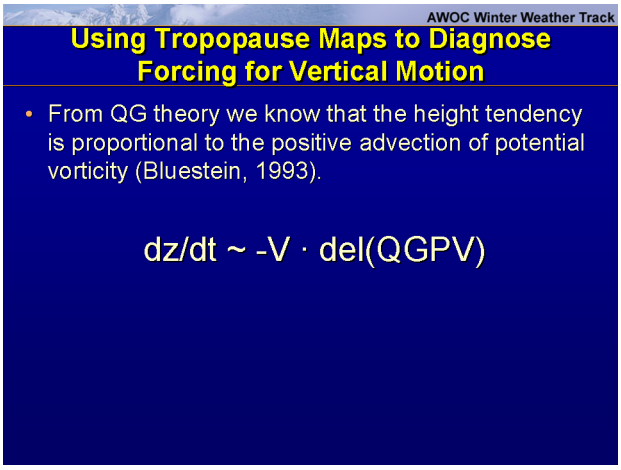
Student Notes:



29. Using Tropopause Maps to Diagnose Forcing for Vertical Motion

Instructor Notes: We have seen how we can diagnose the movements of troughs and ridges as well as jets using tropopause maps. But the question we want to answer when forecasting is how can you use tropopause maps to diagnose forcing for vertical motion. Unlike QG theory, there is no equation that explicitly relates potential vorticity advection to vertical motion. However, one can develop an equation that relates the height tendency to potential vorticity advection. Specifically, height falls are proportional to the amount of positive vorticity advection in the atmosphere. As we discussed earlier, positive vorticity advection at a level (such as 300 mb) similar to positive pressure advection (or negative potential temperature advection) on a tropopause maps. Therefore an area where there is higher pressure being advected on a tropopause map will be an area we would expect height falls. In a couple of slides we will see an example of this with respect to cyclogenesis. Because of the forcing for height rises and falls, a secondary circulation must develop. From the isallobaric equation we know areas of height falls (rises) will be areas where the isallobaric wind will produces convergence (divergence). If this process was the only forcing for vertical motion in the troposphere, then one would produce ascent (descent) where the positive (negative) pressure advection on the tropopause. Let's look at an image.

Student Notes:

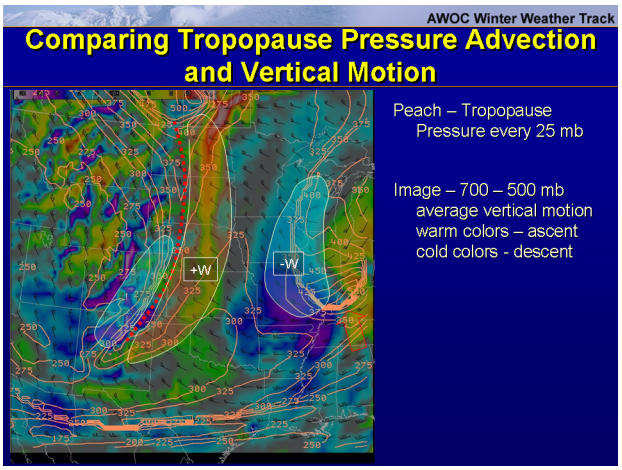


30. Comparing Tropopause Pressure Advection and Vertical Motion

Instructor Notes: Let's look at an example. The first graphic shows two waves. A stronger wave, with the tropopause pressure greater than 500 mb, is located over Indiana and Illinois with negative pressure advection implied across Wisconsin and Illinois. A weaker wave, with tropopause pressures down to 375 mb, extends from North Dakota into New Mexico. The implied positive pressure advection extends across the eastern Dakotas to the Texas panhandle. There is neutral or weakly negative pressure advection over east-

ern Colorado into western New Mexico. Now examine the mean 700-500 mb omega at the same time. We have kept tropopause pressure and wind overlaid to provide context. Looking over the Mississippi River Valley, we see there is generally subsidence over the entire region where there is negative pressure advection. Over the plains, we generally see ascent over the area from the eastern Dakotas into western Texas. However, over the Rockies, there is much less correlation as the influence of terrain overwhelms the influence of the upper level wave. The impact of terrain will be discussed in later lessons. While the correlation is not perfect between the pressure advection on the tropopause and vertical motion, one can say in general that ascent (descent) is associated with areas of positive (negative) pressure advection on the tropopause. As will be discussed in Lessons 2 and 3, the strength of this vertical motion, or how heavy the snow, is determined by mesoscale features, stability, and the degree of interaction between the low-level front and upper-level PV anomaly.

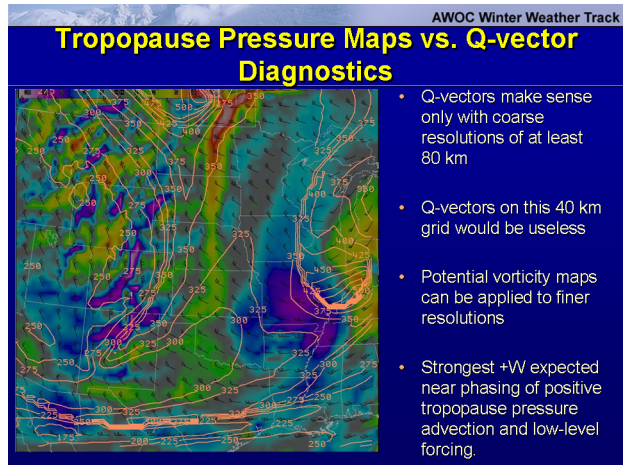
Student Notes:



31. Tropopause Pressure Maps vs. Q-Vector Diagnostics

Instructor Notes: So how is this different from what we already examine, particularly Q-vectors? As noted earlier, we need to use a grid resolution of 80 km or higher. With tropopause maps and PV, there are no such restrictions. We can examine pressure advection on a high resolution grid and determine the large-scale forcing for ascent. This provides a particular advantage when using AWIPS since the higher resolution data has higher time resolution. So we can examine when synoptic-scale forcing for ascent deduced from a tropopause map is collocated with an area with low-level frontogenesis and low stability. In the operational environment, we can use the lower resolution (> 80 km) data to explicitly examine Q-vector convergence for forcing for synoptic-scale ascent and then use pressure advection on tropopause maps on higher resolution (< 80 km) grids when examining mesoscale forcing associated with fronts and the potential interaction between the upper- and lower-level features.

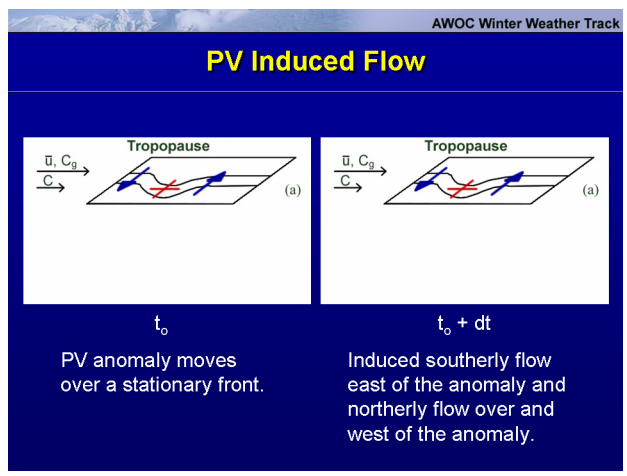
Student Notes:



32. PV Induced Flow

Instructor Notes: As noted earlier, a potential vorticity anomaly can induce flow at low levels. The above illustration shows a tropopause anomaly moving over a stationary front. The PV anomaly itself is associated with a cyclonic rotation. Meanwhile, at the surface, we have no thermal advection with directional shear across the front. However, if the front remains in place, a cyclonic circulation will develop along the surface boundary and a thermal ridge will develop at the surface. Eventually, a cyclone would develop along the front. Notice the cyclonic circulation is displaced downstream from the upper-level PV anomaly.

Student Notes:

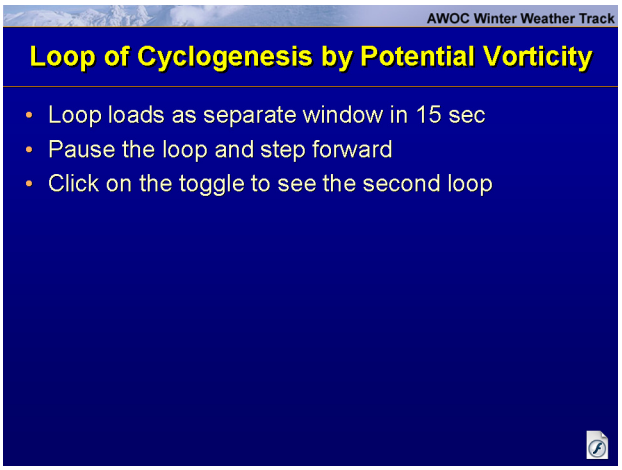


33. Loop of Cyclogenesis by Potential Vorticity

Instructor Notes: Above are two loops showing cyclogenesis from a potential vorticity perspective. In the first loop, above, we examine tropopause pressure, mean sea level pressure, and equivalent potential temperature. Focus on the equivalent potential temperature gradient which extends from western Kansas into Tennessee (and then along the East Coast). An area with higher pressure on the tropopause will move from the

northern Rockies into western Kansas. As the pressure gradient moves over the thermal gradient, a cyclone develops across western Kansas. This cyclone remains in the pressure gradient as it moves across the Mississippi River into Tennessee. However, as the low approaches the Appalachians, a new cyclone develops off the East Coast. Both cyclones move northeast, with the western cyclone weakening and the eastern cyclone strengthening. This occurs despite the fact that the pressure gradient ahead of the western wave is stronger than the gradient and pressure advection to the east. To explain what is happening, we need to examine stability. The second loop shows potential temperature on the tropopause, mean sea level pressure, and the difference between the boundary equivalent potential temperature and potential temperature on the tropopause, which is shaded. The latter is a simple way to assess the stability in the troposphere. When the two are nearly equal (yellow to red shading) then the atmosphere is nearly unstable and the influence of the PV anomaly can more easily reach the surface. As you examine this loop, notice that there is lower stability across western Kansas where the cyclone initially develops. The stability decreases across Tennessee and, as the wave approaches that area, the central pressure does decrease 3-5 mb. However, notice the persistent area of low stability off the southeast coast of the United States. This area of low stability occurs because of the presence of the Gulf Stream results in higher boundary layer equivalent potential temperature. A weak PV anomaly moves from the Gulf of Mexico over this area, just when cyclogenesis commences off the Carolina coast. As the cyclone develops off the coast, the stability increases near the Appalachians and the primary cyclone fills as it moves west of the Appalachians.

Student Notes:

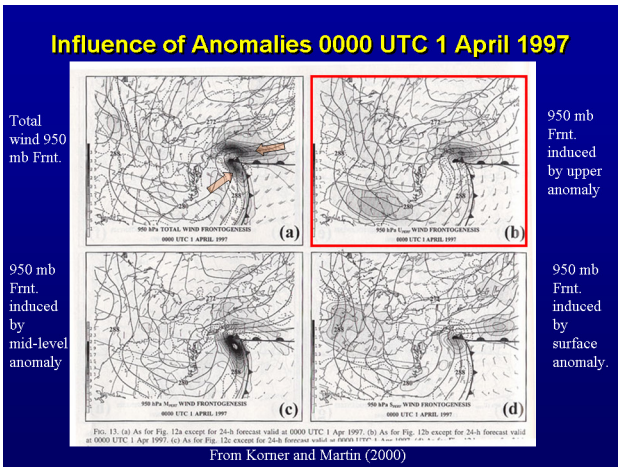


34. Influence of Anomalies: 0000 UTC 1 April 1997

Instructor Notes: Another way to examine the influence of an upper-level wave is to explicitly calculate the low-level wind field induced by the upper-level PV anomaly. Like with Q-vectors, one can partition PV into layers and calculate what the wind field looks like due to that anomaly. Korner and Martin (2000) did so and even calculated frontogenesis from the induced winds. In the upper-left corner is the total frontogenesis. In the upper-right corner, frontogenesis calculated from the 950 mb wind field induced from the upper-level anomaly. What is evident is that a large portion of the frontogenesis along the

warm front is the result of winds induced by the upper-level wave. Being aware of how the upper level wave can influence the winds, such as by increasing deformation, can help forecasters anticipate frontal evolution and circulations.

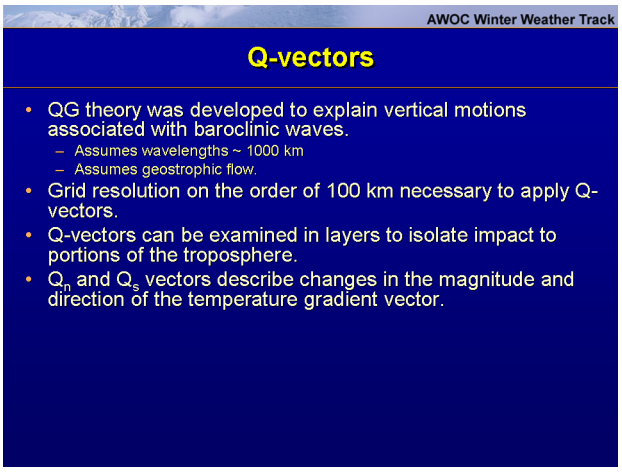
Student Notes:



35. Q-Vectors

Instructor Notes: Listed on the slide is a summary of the topics that were discussed about Q-vectors.

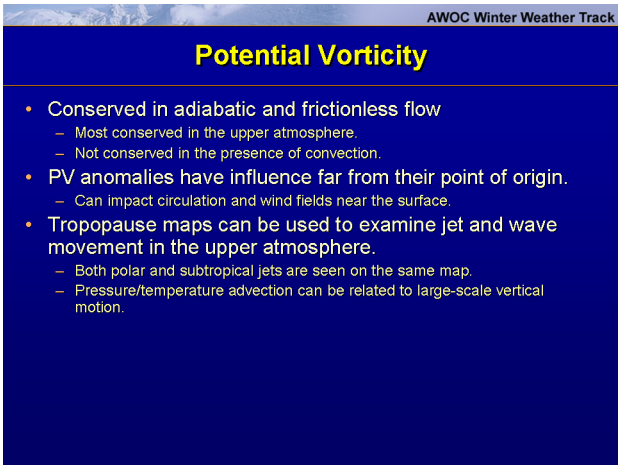
Student Notes:



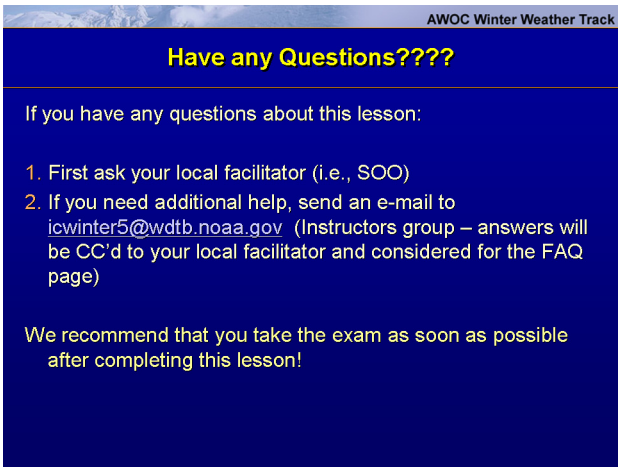
36. Potential Vorticity

Instructor Notes: Listed on the slide is a summary of the topics covered on the subject of Potential Vorticity and tropopause maps.

Student Notes:



37. Have any Questions????

Instructor Notes: If you have any questions about this lesson, first ask your local AWOC facilitator. If you need additional help, send an e-mail to the address provided. When we answer, we will CC your local facilitator and may consider your question for our FAQ page. We strongly recommend that you take the exam as soon as possible after completing this lesson.

Student Notes: