Lecture 6

Wind Veering and Backing, and Geostrophic Temperature Advection

6.1 Introduction

The geostrophic equations:

$$u_g = -\frac{1}{\rho f}\frac{\partial p}{\partial y} \quad and \quad v_g = \frac{1}{\rho f}\frac{\partial p}{\partial x} \quad (6.1)$$

can be written for *isobaric surface* (pressure surface with isohypses) by using the hydrostatic equation ($\frac{\partial p}{\partial z} = -\rho g$) as follows:

$$u_g = -\frac{g}{f}\frac{\partial z}{\partial y} \quad and \quad v_g = \frac{g}{f}\frac{\partial z}{\partial x} \quad (6.2)$$

From the formulas (6.2), it appears that the magnitude of the geostrophic wind depends on the *tilt* of the pressure surface ($\frac{\partial z}{\partial y}$ and $\frac{\partial z}{\partial x}$). For a given distance (∂y or ∂x) of the isohypses the magnitude of the geostrophic wind can be determined with the aid of equation (6.2). In Fig. (6.1), it is obvious that ($\frac{\partial z}{\partial x} < 0$) (height z of the pressure surface decreases in the positive x-direction) and consequently for the v-component of the geostrophic wind we have ($v_g < 0$). The larger the tilt of the pressure surface the smaller the distance (∂x) between the isohypses and hence the larger the geostrophic wind speed. A similar conclusion applies to the u-component of the geostrophic wind. It appears that the magnitude (and the direction) of the geostrophic wind may change if pressure surfaces are NOT parallel to one another.

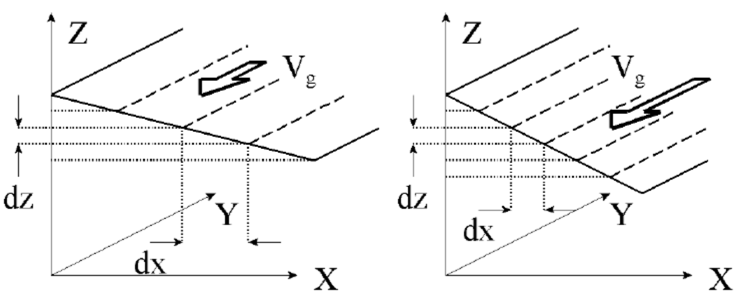


Fig. 6.1 The magnitude of the geostrophic wind is determined by the tilt of the pressure surfaces

6.2 The Relation with Temperature

According to the equation of state (i.e. the gas law) the temperature determines the air density (ρ) for a given value of the pressure (p):

$$p = \rho\, R\, T \quad hence \quad \rho = \frac{p}{R\, T} \qquad (6.3)$$

- In an area with high temperatures, the density at a given pressure is lower than in an area with low temperatures (where the density is higher). Because of the high density, the decrease of pressure with height in the cold area is larger than the decrease of pressure in the warm area.
- In Fig. (6.2) (left) this is clearly visible. We have assumed that the pressure at the surface (z = 0) is the same everywhere (p_0). In the cold area, less vertical distance is needed to have the same decrease in pressure (dp).

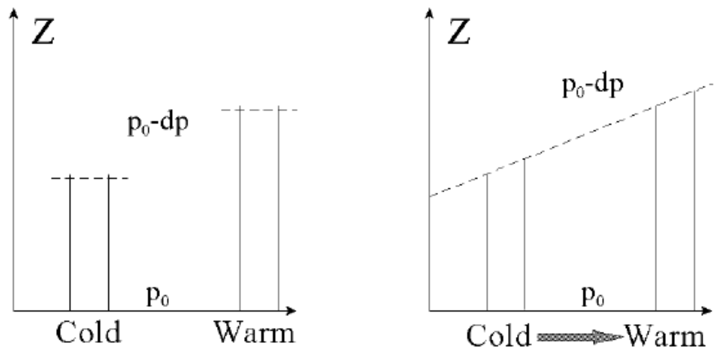


Fig. 6.2. The effect of a horizontal temperature difference (left) and a horizontal temperature gradient (right) on the height and tilt of a pressure surface (assuming equal surface pressure (p_0).

- If temperature changes in the horizontal, then the pressure surface aloft is no longer horizontal but will have a tilt (Fig. 6.2 right). A horizontal temperature gradient influences the tilt of all pressure surfaces.
- From Fig. (6.3), it appears that the tilt of the pressure surfaces increases higher up in the atmosphere. This will have consequences on the magnitude of the geostrophic wind: in the case sketched in Fig. (6.3) it will increase with height. From Figures (6.1 and 6.3), the following conclusion can be drawn:

A horizontal temperature gradient causes a change of the geostrophic wind with height (or with pressure).

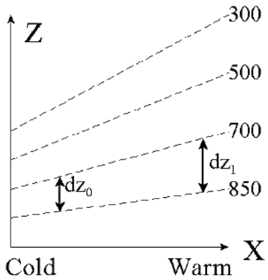


Fig. (6.3.) the effect of a horizontal temperature gradient on the tilt of subsequent pressure surfaces

The difference between the geostrophic wind at the two pressure surfaces is called the thermal wind. An expression for the magnitude of the thermal wind is determined by differentiating the expression of the geostrophic wind with relation to pressure (p):

$$\frac{\partial}{\partial p} u_g = \frac{\partial}{\partial p}\left(-\frac{g}{f}\frac{\partial z}{\partial y}\right) = -\frac{g}{f}\frac{\partial}{\partial p}\left(\frac{\partial z}{\partial y}\right) = -\frac{g}{f}\frac{\partial}{\partial y}\left(\frac{\partial z}{\partial p}\right) \qquad (6.4)$$

For the factor, $\left(\frac{\partial z}{\partial p}\right)$ we use the assumption of hydrostatic equilibrium:

$$\frac{\partial z}{\partial p} = -\frac{1}{\rho g} = -\frac{1}{g}\frac{R\,T}{p}$$

Here we have used the gas law. Substituting this in Equation (6.4) and multiplying by (p) leads to:

$$p\frac{\partial u_g}{\partial p} = \frac{\partial u_g}{\partial \ln p} = \frac{R}{f}\frac{\partial T}{\partial y} \qquad (6.5)$$

We integrate this expression from p_0 to p_1, with $p_0>p_1$:

$$\int_{p_0}^{p_1} \partial u_g = \frac{R}{f}\int_{p_0}^{p_1} \frac{\partial T}{\partial y}\,\partial \ln p$$

This leads to an expression for the components of the **thermal wind** (u_T, v_T):

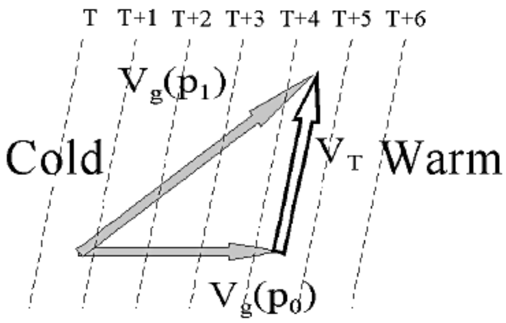
$$u_T = u_g(p_1) - u_g(p_0) = -\frac{R}{f}\frac{\partial \bar{T}}{\partial y}\ln\left(\frac{p_0}{p_1}\right) \qquad (6.6) \qquad and,$$

$$v_T = v_g(p_1) - v_g(p_0) = \frac{R}{f}\frac{\partial \bar{T}}{\partial x}\ln\left(\frac{p_0}{p_1}\right) \qquad (6.7)$$

In the above integration, we have used an average value $\bar{T}$ for the temperature in the layer $p_0 - p_1$ in order to simplify the integration. From these expressions, it is obvious that the thermal wind depends on the horizontal temperature gradient. The following rule also follows from these equations: ***The thermal wind blows parallel to the isotherms with the cold air on the left hand side. The closer the isotherms the stronger the thermal wind.***

6.2 Advection of Warm and Cold Air

From Equations (6.6) and (6.8) it appears that the direction of the thermal wind is parallel to the isotherms. This leads to the following two possible situations.
In Fig. (6.4), the wind is **backing** going from p_0 (low level) to p1<p0 (higher level), in this case from west to southwest. The thermal wind is the wind vector on the high level minus the wind vector on the low level. The thermal wind blows parallel to the isotherms with the cold air on the left hand side. The geostrophic wind on both levels is blowing from the cold area: this is a case of **cold air advection**. In Fig. (6.5), the wind is **veering** from west northwest. The thermal wind again has cold air on its left hand side, and in this case, it appears that the wind on both pressure levels is blowing from the warm area: this is a case of **warm air advection**.



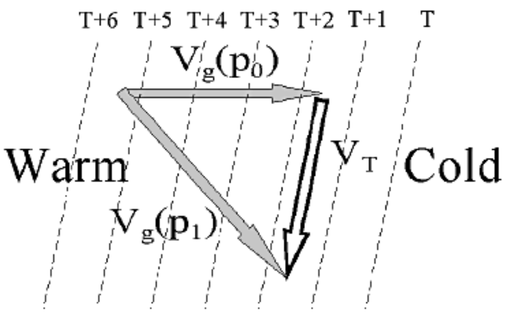


Fig. (6.4) Wind backing with height is an indication of cold air advection

Fig. (6.5) Wind veering with height is an indication of warm air advection

The value of the temperature advection is determined by the value of the thermal wind (i.e. by the distance of the isotherms, i.e. by the horizontal temperature gradient) and by the component of the geostrophic wind which is at right angles to the isotherms.

6.3 Hodograph and Stability

Advection of air from a different average temperature, in a number of layers in the vertical, influences the stability of the atmosphere. It is relatively easy to gain an overview of the advection in different layers by performing what is called a hodograph analysis. This involves the construction of a radial diagram with the station in the center. From this center, the wind speed at several or all pressure levels is plotted as a vector (Fig. 6.6). The vector is plotted in the direction of the wind. This means that a westerly wind is plotted as an arrow to the east (Fig. 6.6). Going to the next higher pressure level, the next arrow is plotted etc. The end points of all vectors, are connected by straight lines (or arrows). These straight lines are the vectors of the thermal wind. The end result is called a hodograph.

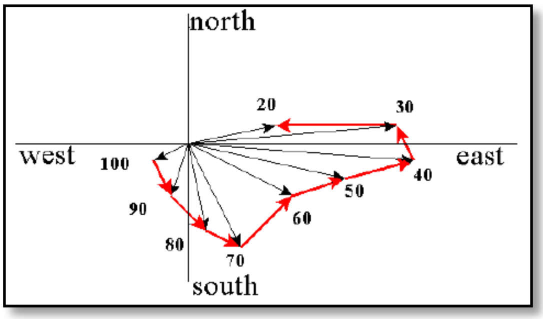


Fig. 6.6. Example of a hodograph

The next step is to consider the veering or backing of the wind, starting from the lowest level to ever higher pressure levels. In this way, the warm or cold air advection in each layer can be determined. By comparing the areas of the triangles in relation to warm or cold air advection, it is possible to determine the changes in atmospheric stability. As an example in Fig. (6.6) between 70 and 50 kPa there is more cold air advection than between 90 and 70 kPa. This means that in this case, the vertical temperature gradient must increase and instability is increasing.