

# CHAPTER 1

## CONVERGENCE, DIVERGENCE, AND VORTICITY

In your reading of the *AG2* manual, volume 1, you became familiar with the terms *convergence*, *divergence*, and *vorticity* when used in relation to surface lows and highs. You were also presented with a basic understanding of the principles involved. In this section, we will cover the terms, the motions involved in upper air features and surface features, and the relationship of these processes to other meteorological applications.

We will first discuss convergence and divergence, followed by a discussion of vorticity.

### NOTE

The World Meteorological Organization adopted "hectopascals" (hPa) as its standard unit of measurement for pressure. Because the units of hectopascals and millibars are interchangeable (1 hPa = 1 mb), hectopascals have been substituted for millibars in this TRAMAN.

### CONVERGENCE AND DIVERGENCE

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**LEARNING OBJECTIVES:** Define the terms *convergence* and *divergence*. Recognize directional and velocity wind shear rules. Recognize areas of mass divergence and mass convergence on surface pressure charts. Identify the isopycnic level. Retail the effects that convergence and divergence have on surface pressure systems and features aloft. Identify rules associated with divergence and convergence.

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As mentioned in the *AG2* manual, volume 1, unit 8, convergence is the accumulation of air in a region or layer of the atmosphere, while divergence is the depletion of air in a region or layer. The layer of maximum convergence and divergence occurs between

the 300- and 200-hPa levels. Coincidentally, this is also the layer of maximum winds in the atmosphere; where jet stream cores are usually found. These high-speed winds are directly related to convergence and divergence. The combined effects of wind direction and wind speed (velocity) is what produces convergent and divergent airflow.

### CONVERGENCE AND DIVERGENCE (SIMPLE MOTIONS)

Simply stated, convergence is defined as the increase of mass within a given layer of the atmosphere, while divergence is the decrease of mass within a given layer of the atmosphere.

#### Convergence

For convergence to take place, the winds must result in a net inflow of air into that layer. We generally associate this type of convergence with low-pressure areas, where convergence of winds toward the center of the low results in an increase of mass into the low and an upward motion. In meteorology, we distinguish between two types of convergence as either horizontal or vertical convergence, depending upon the axis of the flow.

#### Divergence

Winds in this situation produce a net flow of air outward from the layer. We associate this type of divergence with high-pressure cells, where the flow of air is directed outward from the center, causing a downward motion. Divergence, too, is classified as either horizontal or vertical.

### DIRECTIONAL WIND SHEAR

The simplest forms of convergence and divergence are the types that result from wind direction alone. Two flows of air need not be moving in opposite directions to induce divergence, nor moving toward the same point to induce convergence, but maybe at any angle to each other to create a net inflow of air for convergence or a net outflow for divergence.

## WIND SPEED (VELOCITY) SHEAR

Convergence is occurring when wind speeds are decreasing downstream; that is, mass is accumulating upstream. Conversely, divergence is occurring when wind speeds are increasing downstream; that is, mass is being depleted upstream.

## DIRECTIONAL AND SPEED WIND SHEAR

Wind speed in relation to the wind direction is also a valuable indicator. For example, on a streamline analysis chart we can analyze both wind direction and wind speed, variations in wind speed along the streamlines, or the convergence or divergence of the streamlines.

The following are some of the combinations or variations of wind speed and direction:

- In a field of parallel *streamlines* (wind flow), if the wind speed is decreasing downstream (producing a net inflow of air for the layer), convergence is taking place. If the flow is increasing downstream (a net outflow of air from the layer), divergence is occurring.
- In an area of uniform wind speed along the streamlines, if the streamlines diverge (fan out), divergence is occurring; if the streamlines converge (come together), convergence is taking place.
- Normally, the convergence and divergence components are combined. The fact that streamlines converge or diverge does not necessarily indicate convergence or divergence. We must also consider the wind speeds—whether they are increasing or decreasing downstream in relation to whether the streamlines are spreading out or coming together.
- If, when looking downstream on the streamlines, the wind speed increases and the streamlines diverge, divergence is taking place. On the other hand, if the wind speed decreases downstream and the streamlines come together, convergence is taking place.

There are other situations where it is more difficult to determine whether divergence or convergence is occurring, such as when the wind speed decreases downstream and the wind flow diverges, as well as when wind speed increases downstream and the wind flow converges. A special evaluation then must be made to determine the net inflow or outflow.

## DIVERGENCE AND CONVERGENCE (COMPLEX MOTIONS)

In this section we will be discussing high-level convergence and divergence in relation to downstream contour patterns and the associated advection patterns. Low tropospheric advection (and also stratospheric advection) certainly play a large role in pressure change mechanisms.

Since the term *divergence* is meant to denote depletion of mass, while *convergence* is meant to denote accumulation of mass, the forecaster is concerned with the mass divergence or mass convergence in estimating pressure or height changes. Mass divergence in the entire column of air produces pressure or height falls, while mass convergence in the entire column of air produces pressure or height rises at the base of the column.

### Mass Divergence and Mass Convergence

Mass divergence and mass convergence involve the density field as well as the velocity field. However, the mass divergence and mass convergence of the atmosphere are believed to be largely stratified into two layers as follows:

- Below about 600 hPa, velocity divergence and convergence occur chiefly in the friction layer, which is about one-eighth of the weight of the 1,000-to 600-hPa advection stratum, and may be disregarded in comparison with density transport in estimating the contribution to the pressure change by the advection stratum.
- Above 600 hPa, mass divergence and convergence largely result from horizontal divergence and convergence of velocity. However, on occasion, stratospheric advection of density may be a modifying factor.

The stratum below the 400-hPa level may be regarded as the ADVECTION stratum, while the stratum above the 400-mb level maybe regarded as the significant horizontal divergence or convergence stratum. Also, the advection stratum maybe thought of as the zone in which compensation of the dynamic effects of the upper stratum occurs.

### The Isopycnic Level

At about 8km (26,000 ft) the density is nearly constant. This level, which is near the 350-hPa pressure surface, is called the *isopycnic level*. This level is the

location of constant density, with mass variations above and below.

Since the density at 200 hPa is only four-sevenths the density at the isopycnic level, the height change at 200 hPa would have to be twice that at the isopycnic level (350 hPa) for the same pressure/height change to occur. Thus, height changes in the lower stratosphere tend to be a maximum even though pressure changes are a maximum at the isopycnic level.

Pressure changes occur at the isopycnic level, and in order to maintain constant density a corresponding temperature change must also occur. Since the density is nearly constant at this level, the required temperature variations must result from vertical motions. When the pressures are rising at this level, the temperature must also rise to keep the density constant. A temperature rise can be produced by descending motion. Similarly falling pressures at this level require falling temperatures to keep the density constant. Falling temperatures in the absence of advection can be produced by ascent through this level.

Thus, rising heights at the isopycnic level are associated with subsidence, and falling heights at the isopycnic level are associated with convection.

### **The 350-hPa to 200-hPa Stratum**

Subsidence at 350 hPa can result from horizontal convergence above this level, while convection here would result from horizontal divergence above this level.

Since rising heights in the upper troposphere result in a rising of the tropopause and the lower stratosphere, the maximum horizontal convergence must occur between the isopycnic level (350 hPa) and the average level of the tropopause (about 250 hPa). This is due to the reversal of the vertical motion between the tropopause and the isopycnic level. Thus, the level of maximum horizontal velocity convergence must be between 300 hPa and 200 hPa and is the primary mechanism for pressure or height rises in the upper air. Similarly, upper height falls are produced by horizontal velocity divergence with a maximum at the same level. The maximum divergence occurs near or slightly above the tropopause and closer to 200 hPa than to 300 hPa. Therefore, it is more realistic to define a layer of maximum divergence and convergence as occurring between the 300- and 200-hPa pressure surfaces. The 300- to 200-hPa stratum is also the layer in which the core of the jet stream is usually located. It is also at this level that the cumulative effects of the mean temperature

field of the troposphere produce the sharpest horizontal contrasts in the wind field.

The level best suited for determination of convergence and divergence is the 300-hPa level. Because of the sparsity of reports at the 300-hPa level, it is frequently advantageous to determine the presence of convergence and divergence at the 500-hPa level.

### **Divergence/Convergence and Surface Pressure Systems**

The usual distribution of divergence and convergence relative to moving pressure systems is as follows:

- In advance of the low, convergence occurs at low levels and divergence occurs aloft, with the level of nondivergence at about 600 hPa.
- In the rear of the low, there is usually convergence aloft and divergence near the surface.

The low-level convergence ahead of the low occurs usually in the stratum of strongest warm advection, and the low-level divergence in the rear of the low occurs in the stratum of strongest cold advection. The low-level divergence occurs primarily in the friction layer (approximately 3,000 ft) and is thought to be of minor importance in the modification of thickness advection compared with heating and cooling from the underlying surfaces.

### **Divergence/Convergence Features Aloft**

In advance of the low, the air rises in response to the low-level convergence, with the maximum ascending motion at the level of nondivergence eventually becoming zero at the level of maximum horizontal divergence (approximately 300 hPa). Above this level, descending motion is occurring. In the rear of the low, the reverse is true; that is, descending motion in the surface stratum and ascending motion in the upper troposphere above the level of maximum horizontal convergence. In deepening systems, the convergence aloft to the rear of the low is small or may even be negative (divergence). In filling systems, the divergence aloft in advance of the low is small or even negative (convergence).

Thus, in the development and movement of surface highs and surface lows, two vertical circulations are involved, one below and one above the 300-hPa level. The lower vertical circulation is upward in the cyclone

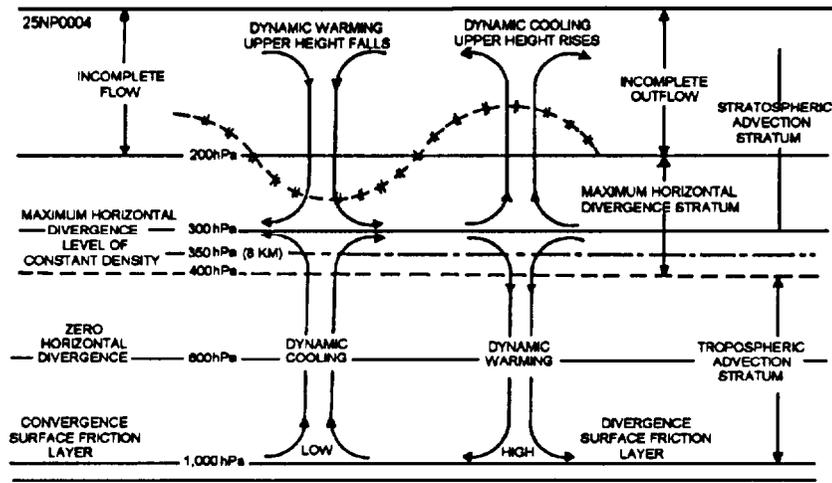


Figure 1-1.-Generalized vertical circulation overdeveloping highs and lows.

and downward in the anticyclone. The upper vertical circulation involves downward motion in the stratosphere of the developing cyclone and upward motion in the upper troposphere and lower stratosphere of the developing anticyclone. See figure 1-1.

Divergence and upper-height falls are associated with high-speed winds approaching weak contour gradients which are cyclonically curved. Figure 1-2 illustrates contour patterns associated with height falls,

Convergence and upper-height rises are associated with the following:

- Low-speed winds approaching straight or cyclonically curved strong contour gradients. See figure 1-3, view (A).
- High-speed winds approaching anticyclonically curved weak contour gradients. See figure 1-3, view (B).

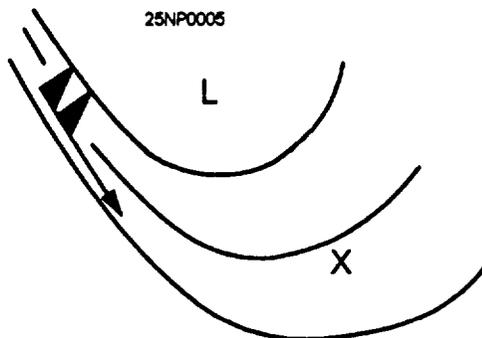


Figure 1-2.-Divergence Illustrated.

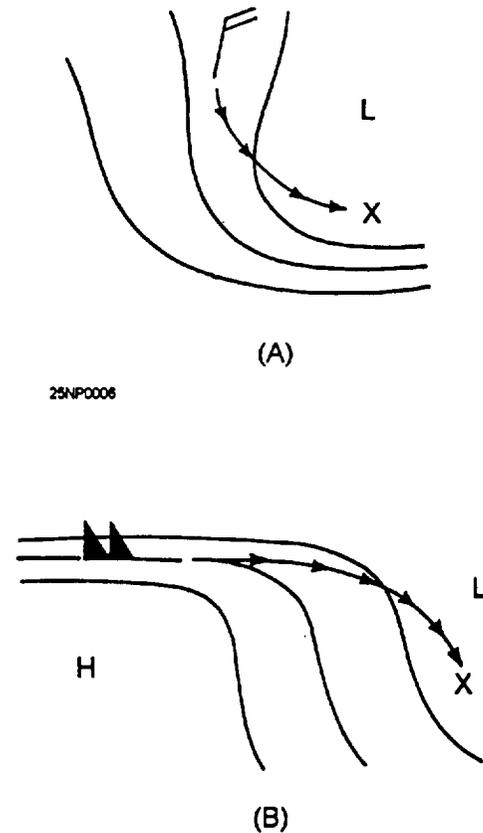


Figure 1-3.-Convergence Illustrated.

Note that the associated height rises or falls occur downstream and to the left of the flow, as illustrated in figure 1-3.

### **Divergence Identification (Downstream Straightline Flow)**

The technique for determining the areas of divergence consists in noting those areas where winds of high speed are approaching weaker downstream gradients that are straight. When inertia carries a high-speed parcel of air into a region of weak gradient, it possesses a Coriolis force too large to be balanced by the weaker gradient force, It is thus deflected to the right. This results in a deficit of mass to the left. The parcels that are deflected to the right must penetrate higher pressure/heights and are thus slowed down until they are in balance with the weaker gradient. Then they can be steered along the existing isobaric or contour channels.

### **Divergence Identification (Weak Downstream Cyclonically Curved Flow)**

If the weak downstream gradients are cyclonically curved, the divergence resulting from the influx of high-speed wind is even more marked due to the additional effect of centrifugal forces.

### **Divergence Identification (Downstream Anticyclonically Curved Flow)**

The effect of centrifugal forces on anticyclonically curving high-speed parcels is of extreme importance in producing overshooting of high-speed air from sharply curved ridges into adjacent troughs, causing pressure rises in the west side of the troughs.

### **Divergence Identification (Strong Winds)**

If high-speed parcels approach diverging cyclonically curved contours, large contour falls will occur downstream to the left of the high-speed winds. Eventually a strong pressure gradient is produced downstream, to the right of the high-speed winds, chiefly as a result of pressure falls to the left of the direction of high-speed winds in the cyclonically curved contours with weak pressure gradient. Usually the deflection of air toward higher pressure is so slight that it is hardly observable in individual wind observations. However, when the pressure field is very weak to the right of the incoming high-speed stream, noticeable angles between the wind and contours may be observed,

especially at lower levels, due to transport of momentum downward as a result of subsidence, where the gradients are even weaker. This occurs sometimes to such an extent that the wind flow is considerably more curved anticyclonically than the contours. In rare cases this results in anticyclonic circulation centers out of phase with the high-pressure center. This is a transitory condition necessitating a migration of the pressure center toward the circulation center. In cases where the high-pressure center and anticyclonic wind flow center are out of phase, the pressure center will migrate toward the circulation center (which is usually a center of mass convergence).

It is more normal, however, for the wind component toward high pressure to be very slight, and unless the winds and contours are drawn with great precision, the deviation goes unnoticed.

### **Overshooting**

High-speed winds approaching sharply curved ridges result in large height rises downstream from the ridge due to overshooting of the high-speed air. It is known from the gradient wind equation that for a given pressure gradient there is a limiting curvature to the trajectory of a parcel of air moving at a given speed. Frequently on upper air charts, sharply curved stationary ridges are observed with winds of high speed approaching the ridge. The existence of a sharply curved extensive ridge usually means a well-developed trough downstream, and frequently a cold or cutoff low exists in this trough. The high-speed winds approaching the ridge, due to centrifugal forces, are unable to make the sharp turn necessary to follow the contours. These winds overshoot the ridge anticyclonically, but with less curvature than the contours, resulting in their plunging across contours toward lower pressure/heights downstream from the ridge. This may result in any one of a number of consequences for the downstream trough, depending on the initial configuration of the ridge and trough, but all of these consequences are based on the convergence of mass into the trough as a result of overshooting of winds from the ridge.

Four effects of overshooting areas follows:

1. Filling of the downstream trough. This happens if the contour gradient is strong on the east side of the trough; that is, a blocking ridge to the east of the trough.
2. Acceleration of the cutoff low from of its stationary position. This usually occurs in all cases.

3. Radical reorientation of the trough. This usually happens where the trough is initially NE-SW, resulting in a N-S and in some cases a NW-SE orientation after sufficient time (36 hours).

4. This situation may actually cut off a low in the lower area of the trough. This usually happens when the high-speed winds approaching the ridge are southwesterly and approach the ridge at a comparatively high latitude relative to the trough. This frequently reorients the trough line towards a more NE-SW direction. Usually, the reorientation of the trough occurs simultaneously with 1 and 2.

### **Sharply Curved Ridges**

Closely related to the previously mentioned situation are cases of sharply curved ridges where the gradient in the sharply curved portion (usually the northern portions of a north-south ridge) has momentarily built up to a strength that is incompatible with the anticyclonic curvature. Such ridges often collapse with great rapidity prior to the development of such excessive gradients, causing rapid filling of the adjacent downstream trough, and large upper contour falls. The gradient wind relation implies that subsequent trajectories of the high-speed parcels generated in the strong ridge line gradient must be less anticyclonically curved than the contours in the ridge.

It can also be shown from the gradient wind equation that the anticyclonic curvature increases as the difference between the actual wind and the geostrophic wind increases, until the actual wind is twice the geostrophic wind, when the trajectory curvature is at a maximum. This fact can be used in determining the trajectory of high-speed parcels approaching sharply curved stationary ridges or sharply curved stationary ridges with strong gradients. By measuring the geostrophic wind in the ridge, the maximum trajectory curvature can be obtained from the gradient wind scale. This trajectory curve is the one that an air parcel at the origin point of the scale will follow until it intersects the correction curve from the geostrophic speed to the displacement curve of twice the geostrophic speed.

### **Actual Wind Speeds**

If actual wind speed observations are available for parcels approaching the ridge, comparison can be made with the geostrophic winds (pressure gradient) in the ridge. If the actual speeds are more than twice the measured geostrophic wind in the ridge, the anticyclonic curvature of these high-speed parcels will be less than

the maximum trajectory curvature obtained from the gradient wind scale, and even greater overshooting of these high-speed parcels will occur across lower contours. Convergence in the west side of the downstream trough results in lifting of the tropopause with dynamic cooling and upper-level contour rises.

### **Subgradient Winds**

Low-speed winds approaching an area of stronger gradient become subject to an unbalanced gradient force toward the left due to the weaker Coriolis force. These subgradient winds are deflected toward lower pressure, crossing contours and producing contour rises in the area of cross-contour flow. This cross-contour flow accelerates the air until it is moving fast enough to be balanced by the stronger pressure gradient. Due to the acceleration of the slower oncoming parcels of air, the contour rises propagate much faster than might be expected on the basis of the slow speed of the air as it initially enters the stronger pressure gradient.

The following two rules summarize the discussion of subgradient winds:

- High-speed winds approaching low-speed winds with weak cyclonically curved contour gradients are indicative of divergence and upper-height falls downstream to the left of the current.
- Low-speed winds approaching strong, cyclonically curved contour gradients or high-speed winds approaching low-speed winds with weak anticyclonically curved contour gradients are indicative of convergence and upper height rises downstream and to the left and right of the current, respectively.

### **IMPORTANCE OF CONVERGENCE AND DIVERGENCE**

Convergence and divergence have a pronounced effect upon the weather occurring in the atmosphere. Vertical motion, either upward or downward, is recognized as an important parameter in the atmosphere. For instance, extensive regions of precipitation associated with extratropical cyclones are regions of large-scale upward motion. Similarly, the nearly cloud-free regions in large anticyclones are regions in which air is subsiding. Vertical motions also affect temperature, humidity, and other meteorological elements.

## Changes in Stability

When convergence or divergence occurs, whether on a large or small scale, it may have a very pronounced effect on the stability of the air. For example, when convection is induced by convergence, air is forced to rise without the addition of heat. If this air is unsaturated, it cools first at the dry adiabatic rate; or if saturated at the moist rate. The end result is that the air is cooled, which will increase the instability of that air column due to a net release of heat. Clouds and weather often result from this process.

Conversely, if air subsides, and this process is produced by convergence or divergence, the sinking air will heat at the dry adiabatic lapse rate due to compression. The warming at the top of an air column will increase the stability of that air column by reducing the lapse rate. Such warming often dissipates existing clouds or prevents the formation of new clouds. If sufficient warming due to the downward motion takes place, a subsidence inversion is produced.

## Effect on Weather

The most important application of vertical motion is the prediction of rainfall probability and rainfall amount. In addition, vertical motion affects practically all meteorological properties, such as temperature, humidity, wind distribution, and particularly stability. In the following section the distribution of large-scale and small-scale vertical motions are considered.

Since cold air has a tendency to sink, subsidence is likely to be found to the west of upper tropospheric troughs, and rising air to the east of the troughs. Thus, there is a good relation between upper air meridional flow and vertical flow.

In the neighborhood of a straight Northern Hemisphere jet stream, convergence is found to the north of the stream behind centers of maximum speed as well as to the south and ahead of such centers. Divergence exists in the other two quadrants. Below the regions of divergence the air rises; below those of convergence there is subsidence.

These general rules of thumb are not perfect, and only yield a very crude idea about distribution of vertical motion in the horizontal. Particularly over land in summer, there exists little relation between large-scale weather patterns and vertical motion. Rather, vertical motion is influenced by local features and shows strong diurnal variations. Large-scale vertical motion is of small magnitude at the ground (zero if the ground is flat).

Above ground level, it increases in magnitude to at least 500 hPa and decreases in the neighborhood of the tropopause. There have been several studies of the relation between frontal precipitation and large-scale vertical velocities, computed by various techniques. In all cases, the probability of precipitation is considerably higher in the 6 hours following an updraft than following subsidence. Clear skies are most likely with downdrafts. On the other hand, it is not obvious that large-scale vertical motion is related to showers and thunderstorms caused during the daytime by heating. However, squall lines, which are formed along lines of horizontal convergence, show that large-scale vertical motion may also play an important part in convective precipitation.

## Vertical Velocity Charts

Vertical velocity charts are currently being transmitted over the facsimile network and are computed by numerical weather prediction methods. The charts have plus signs indicating upward motion and minus signs indicating downward motion. The figures indicate vertical velocity in centimeters per second (cm/sec). With the larger values of upward motions (plus values) the likelihood of clouds and precipitation increases. However, an evaluation of the moisture and vertical velocity should be made to get optimum results. Obviously, upward motion in dry air is not as likely to produce precipitation as upward motion in moist air.

Studies have shown that surface cyclones and anticyclones are not independent of developments in the upper atmosphere, rather, they work in tandem with one another. The relationship of the cyclone to the large-scale flow patterns aloft must therefore be a part of the daily forecast routine.

Many forecasters have a tendency to shy away from the subject of vorticity, as they consider it too complex a subject to be mastered. By not considering vorticity and its effects, the forecaster is neglecting an important forecasting tool. The principles of vorticity are no more complicated than most of the principles of physics, and can be understood just as readily. In the following section we will discuss the definition of vorticity, its evaluation, and its relationships to other meteorological parameters.

## VORTICITY

**LEARNING OBJECTIVES:** Recognize the two components of relative vorticity. Define the term *absolute vorticity*. Determine vorticity impacts on weather processes.

Vorticity measures the rotation of very small air parcels. A parcel has vorticity when it spins on its axis as it moves along its path. A parcel that does not spin on its axis is said to have zero vorticity. The axis of spinning or rotation can extend in any direction, but for our purposes, we are mainly concerned with the rotational motion about an axis that is perpendicular to the surface of the Earth. For example, we could drop a chip of wood into a creek and watch its progress. The chip will move downstream with the flow of water, but it may or may not spin as it moves downstream. If it does spin, the chip has vorticity. When we try to isolate the cause of the spin, we find that two properties of the flow of water cause the chip to spin: (1) If the flow of water is moving faster on one side of the chip than the other, this is shear of the current; (2) if the creek bed curves, the path has curvature. Vorticity always applies to extremely small air parcels; thus, a point on one of our upper air charts may represent such a parcel. We can examine this point and say that the parcel does or does not have vorticity. However, for this discussion, larger parcels will have to be used to more easily visualize the effects. Actually, a parcel in the atmosphere has *three* rotational motions at the same time: (1) rotation of the parcel about its own axis (*shear*), (2) rotation of the parcel about the axis of a pressure system (*curvature*), and (3) rotation of the parcel due to the atmospheric rotation. The sum of the first two components is known as *relative vorticity*, and the sum total of all three is known as *absolute vorticity*.

### RELATIVE VORTICITY

Relative vorticity is the sum of the rotation of the parcel about the axis of the pressure system (curvature) and the rotation of the parcel about its own axis (shear).

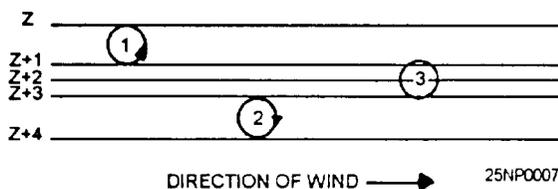


Figure 1-4. Illustration of vorticity due to the shear effect.

The vorticity of a horizontal current can be broken down into two components, one due to curvature of the streamlines and the other due to shear in the current.

### Shear

First, let us examine the shear effect by looking at small air parcels in an upper air pattern of straight contours. Here the wind shear results in each of the three parcels having different rotations (fig. 1-4).

Refer to figure 1-4. Parcel No. 1 has stronger wind speeds to its right. As the parcel moves along, it will be rotated in a counterclockwise direction. Parcel No. 2 has the stronger wind speeds to its left; therefore, it will rotate in a clockwise direction as it moves along. Parcel No. 3 has equal wind speeds to the right and left. It will move, but it will not rotate. It is said to have zero vorticity.

Therefore, to briefly review the effect of shear—a parcel of the atmosphere has vorticity (rotation) when the wind speed is stronger on one side of the parcel than on the other.

Now let's define positive and negative vorticity in terms of clockwise and counterclockwise rotation of a parcel. The vorticity is positive when the parcel has a counterclockwise rotation (cyclonic, Northern Hemisphere) and the vorticity is negative when the parcel has clockwise rotation (anticyclonic, Northern Hemisphere).

Thus, in figure 1-4, parcel No. 1 has positive vorticity, and parcel No. 2 has negative vorticity.

### Curvature

Vorticity can also result due to curvature of the airflow or path. In the case of the wood chip flowing with the stream, the chip will spin or rotate as it moves along if the creek curves.

To demonstrate the effect of curvature, let us consider a pattern of contours having curvature but no shear (fig. 1-5).

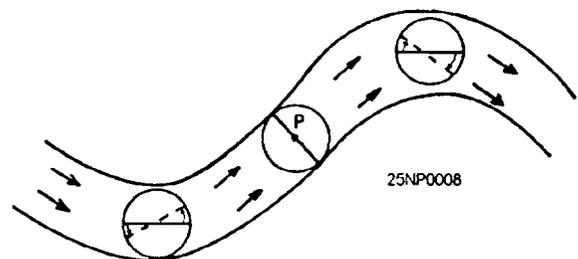


Figure 1-5. Illustration of vorticity due to curvature effect.

Place a small parcel at the trough and ridge lines and observe the way the flow will spin the parcel, causing vorticity. The diameter of the parcel will be rotated from the solid line to the dotted position (due to the northerly and southerly components of the flow on either side of the trough and ridge lines).

Note that we have counterclockwise rotation at the trough (positive vorticity), and at the ridge line we have clockwise rotation (negative vorticity). At the point where there is no curvature (inflection point), there is no turning of the parcel, hence no vorticity. This is demonstrated at point P in figure 1-5.

### Combined Effects

To find the relative vorticity of a given parcel, we must consider both the shear and curvature effects. It is quite possible to have two effects counteract each other; that is, where shear indicates positive vorticity but curvature indicates negative vorticity, or vice versa (fig. 1-6).

To find the net result of the two effects we would measure the value of each and add them algebraically. The measurement of vorticity will be discussed in the next section.

It must be emphasized here that relative vorticity is observed instantaneously. Relative vorticity in the atmosphere is defined as the instantaneous rotation of very small particles. The rotation results from wind shear and curvature. We refer to this vorticity as being relative, because all the motion illustrated was relative to the surface of the Earth.

### ABSOLUTE VORTICITY

When the relative vorticity of a parcel of air is observed by a person completely removed from the Earth, he or she observes an additional component of vorticity created by the rotation of the Earth. Thus, this

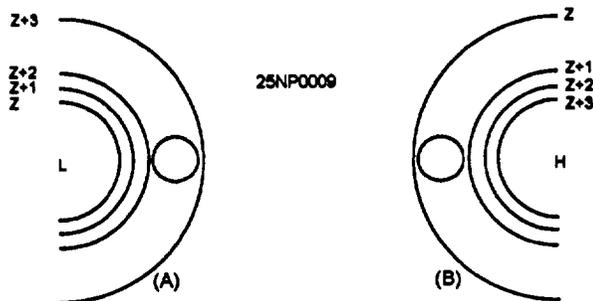


Figure 1-6-Illustration of shear effect opposing the curvature effect in producing vorticity. (A) Negative shear and positive curvature; (B) positive shear and negative curvature.

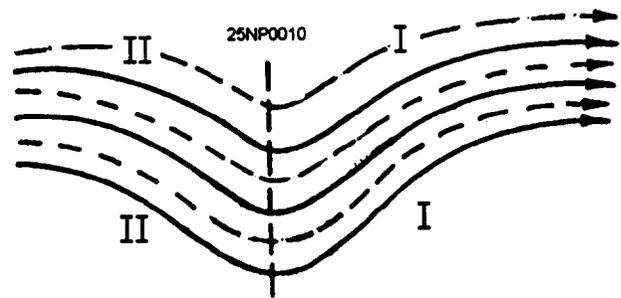


Figure 1-7-Contour-isotach pattern for shear analysis.

person sees the total or absolute vorticity of the same parcel of air.

The total vorticity, that is, relative vorticity plus that due to the Earth's rotation, is known as the absolute vorticity. As was stated before, for practical use in meteorology, only the vorticity about an axis perpendicular to the surface of the Earth is considered. In this case, the vorticity due to the Earth's rotation becomes equal to the Coriolis parameter. This is expressed as  $2\omega \sin \theta$ , where  $\omega$  is the angular velocity of the Earth and  $\theta$  is the latitude. Therefore, the absolute vorticity is equal to the Coriolis parameter plus the relative vorticity. Writing this in equation form gives: (Za = absolute vorticity)

$$Za = 2\omega \sin \theta + Zr$$

### EVALUATION OF VORTICITY

In addition to locating the areas of convergence and divergence, we must also consider the effects of horizontal wind shear as it affects the relative vorticity, and hence the movement of the long waves and deepening or falling associated with this movement.

The two terms *curvature* and *shear*, which determine the relative vorticity, may vary inversely to each other. Therefore, it is necessary to evaluate both of them. Figures 1-7 through 1-10 illustrate some of the possible combinations of curvature and shear. Solid

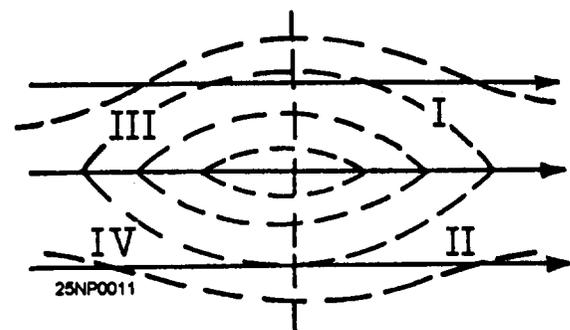


Figure 1-8-Contour-isotach pattern for shear analysis.

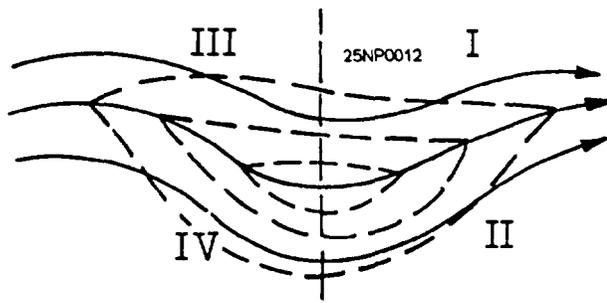


Figure 1-9.-Contour-isotach pattern for shear analysis.

lines are streamlines or contours; dashed lines are isotachs.

Figure 1-7 represents a symmetrical sinusoidal streamline pattern with isotachs parallel to contours. Therefore, there is no gradient of shear along the contours. In region I, the curvature becomes more anticyclonic downstream, reaching a maximum at the axis of the downstream ridge; that is, relative vorticity decreases from the trough to a minimum at the downstream ridge. The region from the trough to the downstream ridge axis is favorable for deepening.

The reverse is true west of the trough, region II. This region is unfavorable for deepening.

In figure 1-8 there is no curvature of streamlines; therefore, the shear alone determines the relative vorticity. The shear downstream in regions I and IV becomes less cyclonic; in regions II and III, it becomes more cyclonic. Regions I and IV are therefore favorable for deepening downstream.

In region I of figure 1-9 both cyclonic shear and curvature decrease downstream and this region is highly favorable for deepening. In region III both cyclonic shear and curvature increase downstream and this region is unfavorable for deepening. In region II the cyclonic curvature decreases downstream, but the cyclonic shear increases. This situation is indeterminate without calculation unless one term predominates. If the curvature gradient is large and the shear gradient small, the region is likely to be favorable for deepening.

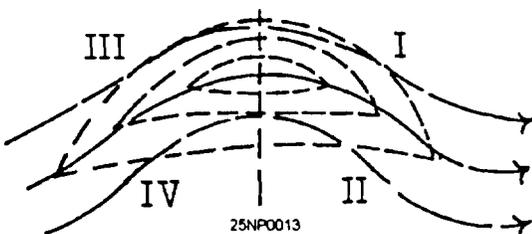


Figure 1-10.-Contour-isotach pattern for shear analysis.

In region IV, the cyclonic curvature increases downstream, but the cyclonic shear decreases, so that this region is also indeterminate unless one of the two terms predominates.

In region I of figure 1-10 the cyclonic shear decreases downstream and the cyclonic curvature increases. The region is indeterminate; however, if the shear gradient is larger than the curvature gradient, deepening is favored. Region II has increasing cyclonic shear and curvature downstream and is quite unfavorable. In region III, the shear becomes more cyclonic downstream and the curvature becomes less cyclonic. This region is also indeterminate unless the curvature term predominates. In region IV, the shear and curvature become less cyclonic downstream and the region is favorable for deepening.

## RELATION OF VORTICITY TO WEATHER PROCESSES

Vorticity not only affects the formation of cyclones and anticyclones, but it also has a direct bearing on cloudiness, precipitation, pressure, and height changes. Vorticity is used primarily in forecasting cloudiness and precipitation over an extensive area. One rule states that when relative vorticity decreases downstream in the upper troposphere, convergence is taking place in the lower levels. When convergence takes place, cloudiness and possibly precipitation will prevail if sufficient moisture is present.

One rule using vorticity in relation to cyclone development stems from the observation that when cyclone development occurs, the location, almost without exception, is in advance of an upper trough. Thus, when an upper level trough with positive vorticity advection in advance of it overtakes a frontal system in the lower troposphere, there is a distinct possibility of cyclone development at the surface. This is usually accompanied by deepening of the surface system. Also, the development of cyclones at sea level takes place when and where an area of positive vorticity advection situated in the upper troposphere overlies a slow moving or quasi-stationary front at the surface.

The relationship between convergence and divergence can best be illustrated by the term shear. If we consider a flow where the cyclonic shear is decreasing downstream (stronger wind to the right than to the left of the current), more air is being removed from the area than is being fed into it, hence a net depletion of mass aloft, or divergence. Divergence aloft is associated with surface pressure falls, and since this is

the situation, the relative vorticity is decreasing downstream. We may state that surface pressure falls where relative vorticity decreases downstream in the upper troposphere, or where advection of more cyclonic vorticity takes place aloft. The converse of this is in the case of convergence aloft.

### **SUMMARY**

In this chapter we expanded on the subjects of convergence, divergence, and vorticity, which were first presented in the AG2 manual, volume 1. Our discussion

first dealt with convergence and divergence as simple motions. The dynamics of convergent and divergent flow was covered, along with a discussion of wind directional shear and wind speed shear. Convergence and divergence as complex motions were then presented. Rules of thumb on convergence and divergence relative to surface and upper air features were covered. The last portion of the chapter dealt with vorticity. Definitions of relative vorticity and absolute vorticity were covered. Vorticity effects on weather processes was the last topic of discussion.

