In a developing science, theoretical insight and observational fact are seldom in balance, and then only momentarily. Since 1940, developments in aircraft, instrumentation, rocketry, and satellites have given meteorologists access to coordinated observations of sufficient scope and detail to permit study of the atmosphere as a global system. At the moment we cannot *explain* these observations theoretically; however, we can *interpret* them to demonstrate that the internal mechanism of the general circulation is dynamically consistent in certain important respects. This chapter will be devoted to that demonstration.

8-2 OBSERVATIONAL DESCRIPTION OF THE GENERAL CIRCULATION[1]

Pressure distribution

The observed distribution of surface-pressure intensity is shown in Fig. 8-1. In this illustration, long-term averages for the months of January and July are presented in millibars corrected to sea level. The principal feature of these distributions is the occurrence of centers of high pressure (marked *H*) at about 30° latitude in both hemispheres. Low-pressure centers (marked *L*) are also found near the Equator in both hemispheres and in the high latitudes, above 45°, in the Northern Hemisphere. These tendencies are demonstrated clearly in Fig. 8-2, where the surface pressures of Fig. 8-1 are presented as zonal averages.[2] Notice that the averages for the ocean only and for the whole earth are almost identical in the Southern Hemisphere, where oceans predominate. The differences between these curves in the Northern Hemisphere reflect the relative temperatures of land and water (see Fig. 6-5).

Velocity distribution

Figure 8-3 shows the streamlines of the mean observed surface wind[3] in July and January. Notice the large *centers* of diverging (called *anticyclonic*) flow and their coincidence with the centers of high pressure

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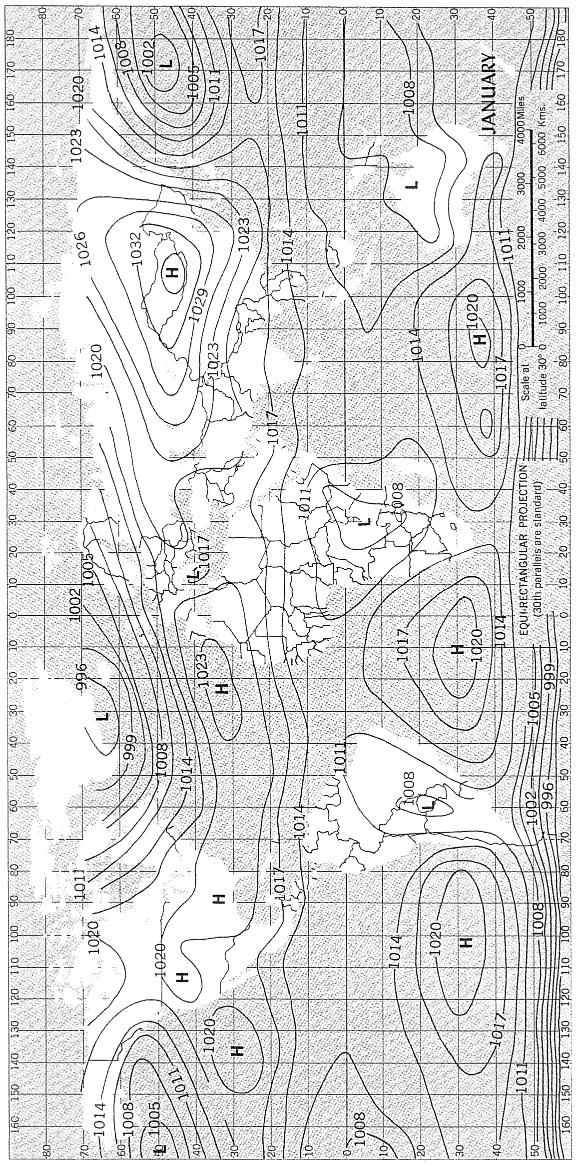
[1] The material of this section is taken largely from the very thorough study by Mintz and Dean [1].

[2] The term *zonal average* refers to averages taken over all longitudes at a given latitude. The distribution, with latitude, of the zonal averages (or some other quantity) is called a *meridional distribution*. East-west components of vector quantities are known as *zonal components*, and north-south components are called *meridional components*.

[3] *Surface* wind refers to that measured at a standard elevation, a few feet above the earth's surface.

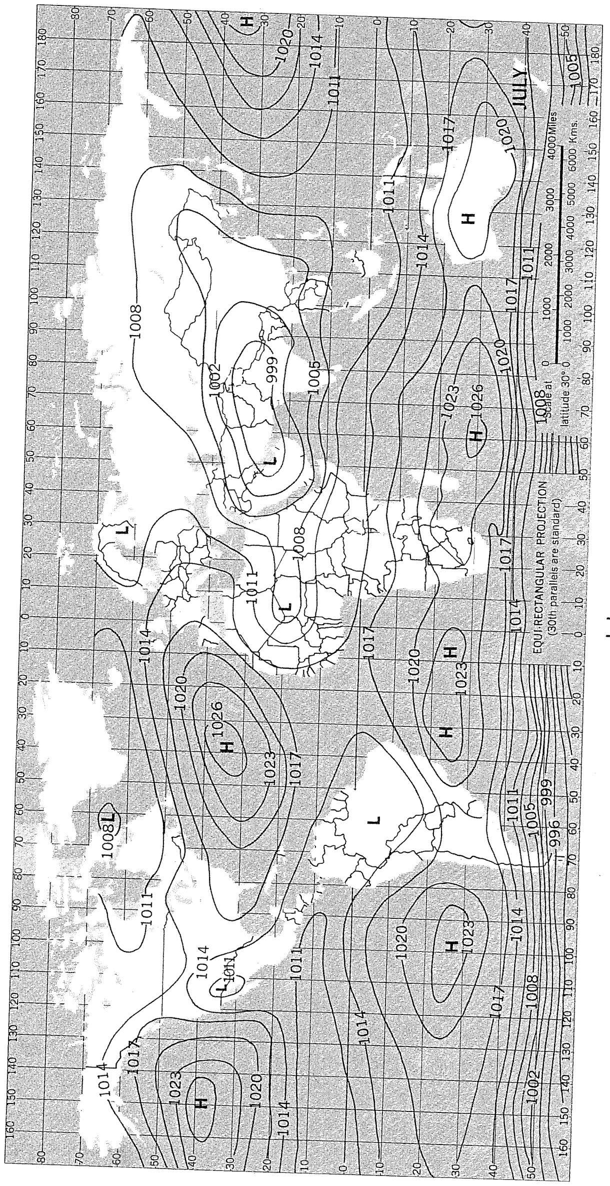
ADDITIONAL BACKGROUND READING



January

(*a*)

Fig. 8-1 Distribution of mean sea-level pressure intensity, mb. (*By permission from G. T. Trewartha, "An Introduction to Climate," 4th ed., McGraw-Hill Book Company, New York, 1968.*)



July

(*b*)

Fig. 8-1 (*Continued.*)

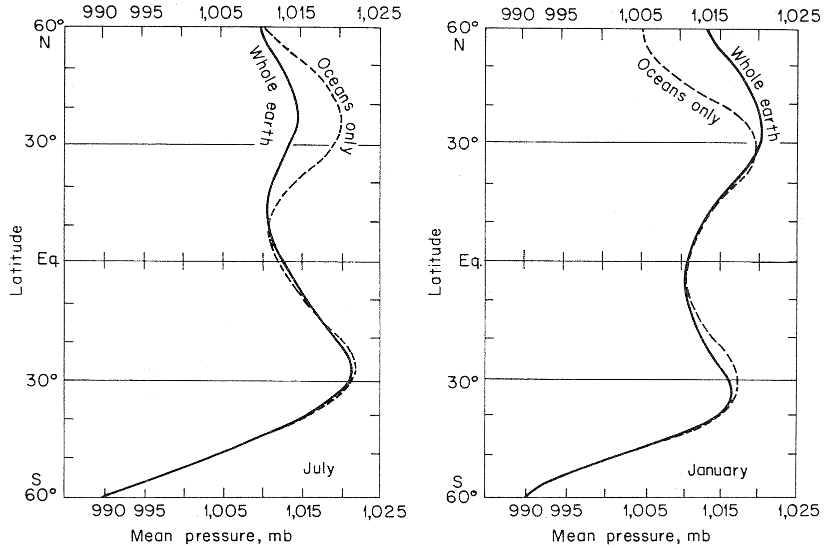
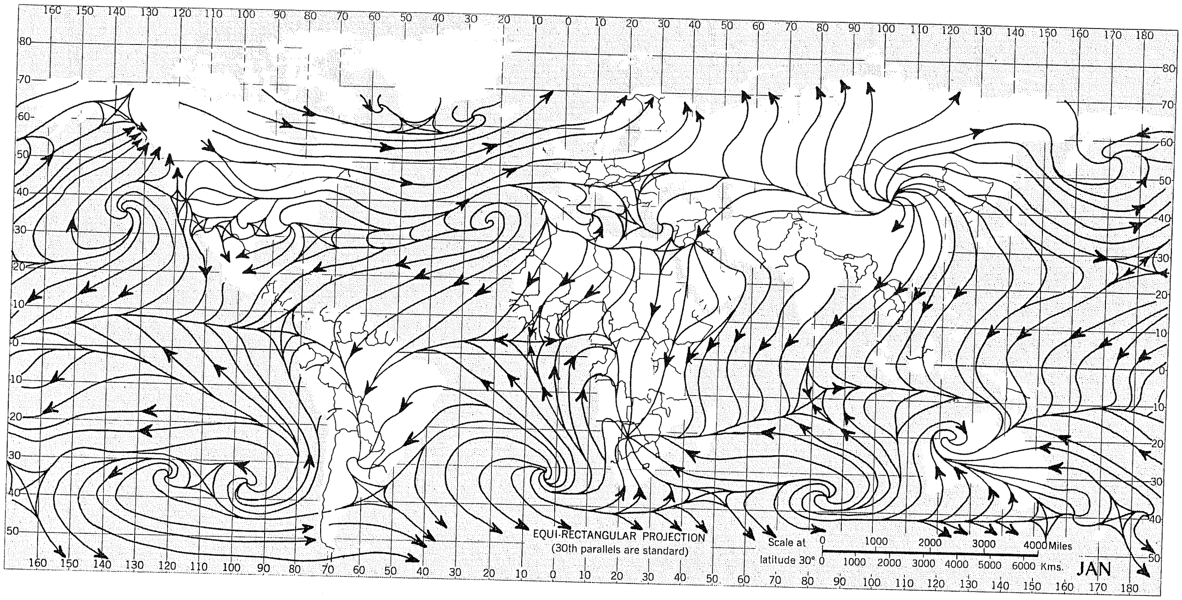


Fig. 8-2 Meridional distribution of average sea-level pressure. (*From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper, 17, Air Force Cambridge Research Center, Cambridge, Mass.,* 1952.)

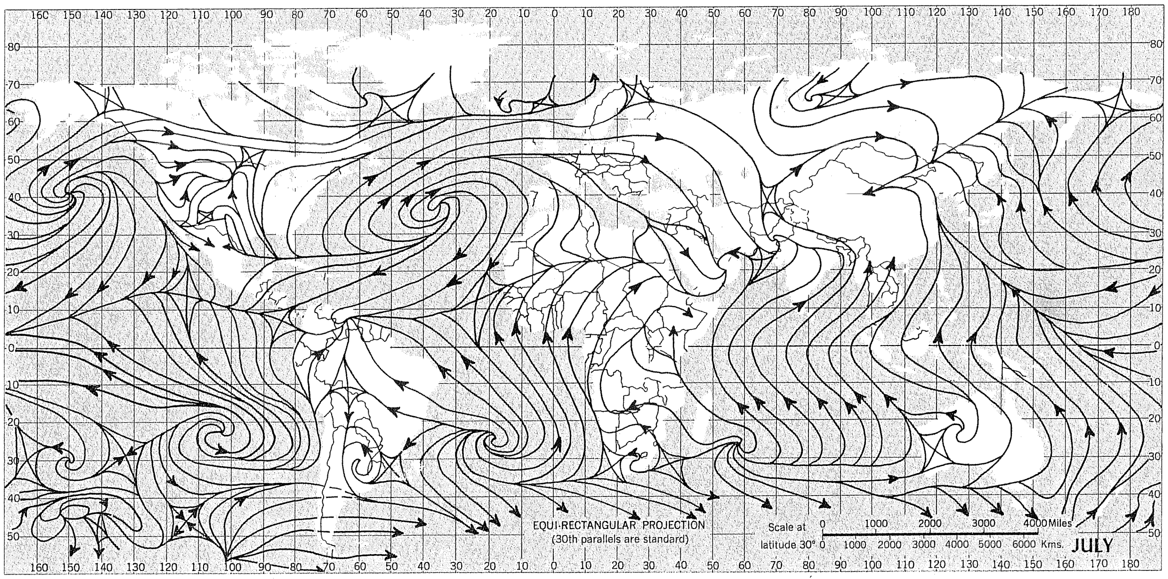
indicated on Fig. 8-1. Although these systems dominate the mean maps, similar but smaller converging surface-wind systems (called *cyclones*) can be seen to correspond with the centers of low pressure. If we examine the anticyclones, they can be seen to move eastward in the summer (January for the Southern Hemisphere, July in the Northern Hemisphere) and, in the Northern Hemisphere at least, to grow in size during the summer.

Another prominent feature of these mean surface winds is the pattern of *asymptotes* of divergence and convergence of wind direction. Predominant are the lines of divergence which are directed, from the centers of divergence, westward toward the Equator and eastward toward the poles. The lines of convergence leading to the cyclones are not as prominent, with the exception of the long line of intertropical convergence which nearly circles the earth in the low latitudes, lying in the equatorial trough of low mean pressure between the anticyclones of the Northern and Southern Hemispheres.

Perhaps the best-known feature of the mean surface wind is the geographical distribution of its zonal component. This is shown in Fig. 8-4. The dark lines separate the zones of tropical easterlies (called *trade winds*) from those of extratropical westerlies. The average of this zonal component over all longitudes is shown as a function of latitude in Fig. 8-5*a*.



(a)



(b)

Fig. 8-3 Observed direction of the mean surface wind. (*a*) January; (*b*) July. (*After Mintz and Dean. By permission from G. T. Trewartha, "An Introduction to Climate," 4th ed., McGraw-Hill Book Company, New York,* 1968.)

Wind observations at higher altitudes are not as complete; however, at levels outside the viscous boundary layer the velocity (called the *geostrophic* velocity) may be calculated quite accurately from the more plentiful observations of pressure and temperature. The 500-mb surface is representative of upper-atmosphere velocities, since it is approximately in the middle of the troposphere with respect to elevation and in the middle of the entire atmosphere as far as mass is concerned. Streamlines and isovels of the mean wind at the 500-mb level in the Northern Hemisphere are shown in Figs. 8-6 and 8-7 for the months of January and July, respectively. Note from these figures that in the middle and higher latitudes, the mean velocity has the appearance of a westerly vortex centered on the earth's rotational axis. The streamlines of this vortex have an irregular waveshape at the higher latitudes. The waves have troughs in eastern North America, in the western Pacific, and in eastern Europe, but they are all damped out as the latitude of maximum velocity is approached.

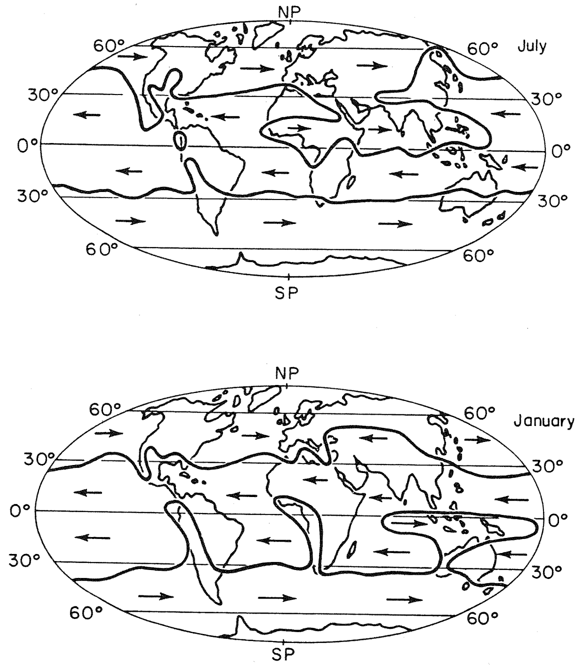


Fig. 8-4 Geographical distribution of the zonal component of the mean surface wind. (*From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper,* 17, *Air Force Cambridge Research Center, Cambridge, Mass.,* 1952.)

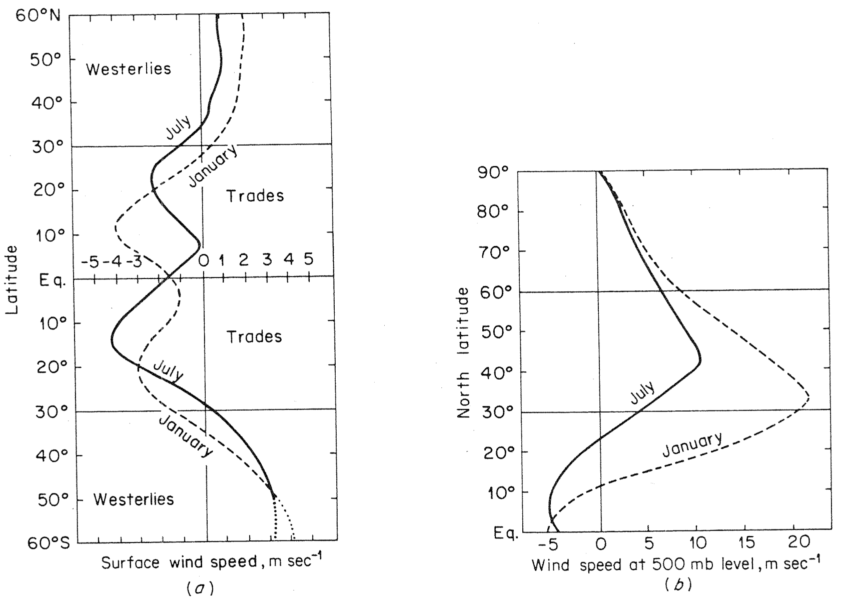


Fig. 8-5 Average zonal component of the observed mean wind. (*a*) Surface; (*b*) 500-mb level. (*From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper,* 17, *Air Force Cambridge Research Center, Cambridge, Mass.,* 1952.)

This latitude varies from about 32°N in the winter to about 45°N in the summer, as shown by the zonal averages in Fig. 8-5*b*. In comparing these velocities with the surface velocities (Fig. 8-5*a*) we see that the upper-air velocities are an order of magnitude larger and that the westerly components predominate. A further view of the concentration of these upper-air westerlies is given by the atmospheric cross sections of Fig. 8-8. The zone of maximum velocity is often referred to as the *jet stream*. Note that it varies in both magnitude and location with season, intensifying and shifting to lower latitudes in the winter. Note also the steep velocity gradients over the vertical at altitudes well outside the frictional boundary layer. Returning now to Figs. 8-6 and 8-7, we see that the transition from the middle-latitude westerlies to the relatively weak, lower-latitude easterlies is through a number of anticyclonic cells located at about 23°N in July and 12°N in January.

Humidity distribution

Because of the importance of water vapor to atmospheric energetics, it is useful to examine the distribution of precipitable water W in the atmosphere. If we assume a hydrostatic atmosphere, the instantaneous precipitable water contained in a unit column of air above any point is given by

$$W = \int_{\infty}^{0} \rho q_h \, dz \approx \frac{1}{g} \int_{0}^{p_0} q_h \, dp \qquad (8\text{-}1)$$

where p_0 = (absolute) surface-pressure intensity
q_h = specific humidity

As a contribution to the International Geophysical Year, Starr et al. [2] used aerological data to calculate time averages $\bar{W}$ of precipitable water for the year 1958 over the Northern Hemisphere. These are shown in Fig. 8-9, where centers of high and low moisture content are indicated by H and L, respectively. Notice that $\bar{W}$ decreases continuously from the Equator to the North Pole (this is clarified by the zonal averages in Fig. 8-10) and that continents and oceans show contrasting influences. The large desert areas in North Africa, the Middle East, and north of Tibet have low values of $\bar{W}$, as do the areas of high elevation in the western United States, central Mexico, the Himalayas, Tibet, central Asia, and central Africa. The western parts of the subtropical oceanic anticyclones have a noticeably higher water content than the eastern parts. Oceanic centers of high precipitable water bracket India and lie at both the eastern and the western equatorial limits of the Pacific Ocean. Continental centers of high $\bar{W}$ are in the equatorial regions of South America and of West Africa.

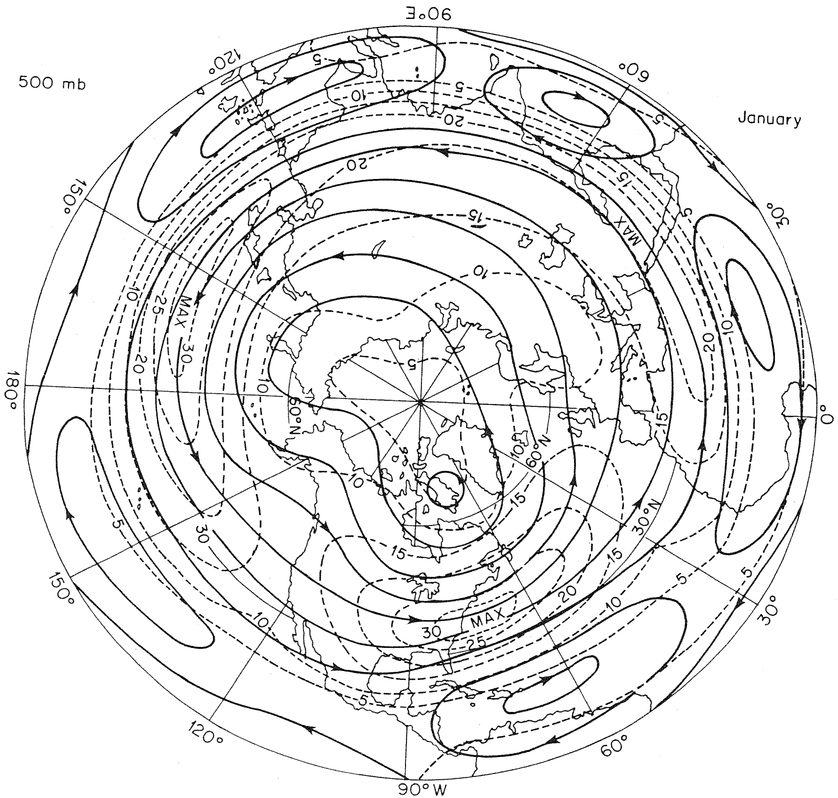


Fig. 8-6 Streamlines and isovels of the Northern Hemisphere mean wind at the 500-mb level in January (speed in meters per second). *(From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper, 17, Air Force Cambridge Research Center, Cambridge, Mass., 1952.)*

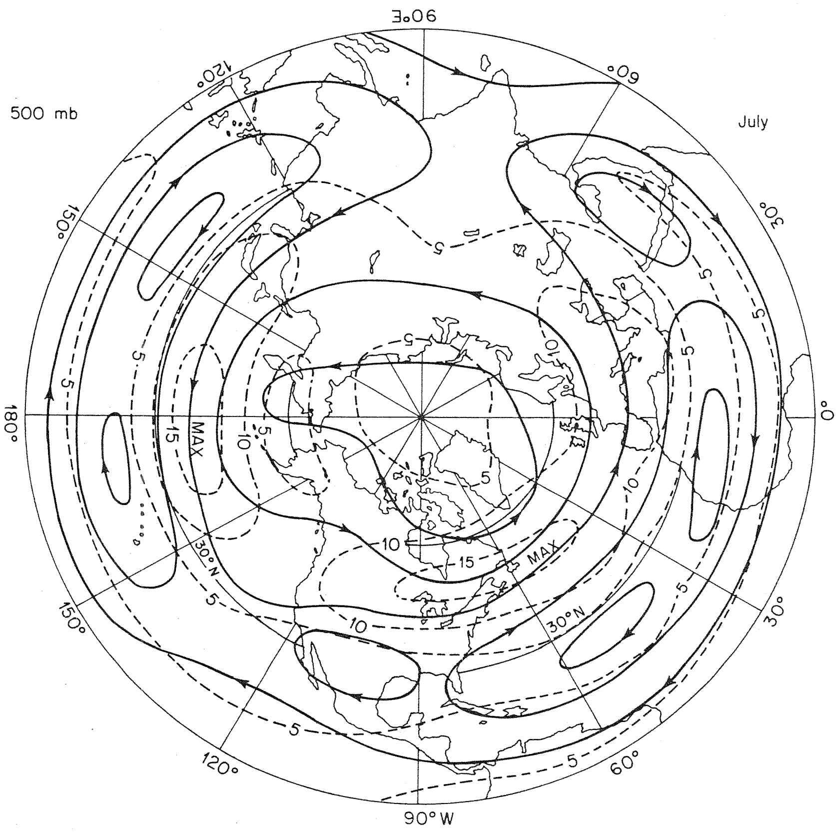


Fig. 8-7 Streamlines and isovels of the Northern Hemisphere mean wind at the 500-mb level in July (speed in meters per second). *(From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper, 17, Air Force Cambridge Research Center, Cambridge, Mass., 1952.)*

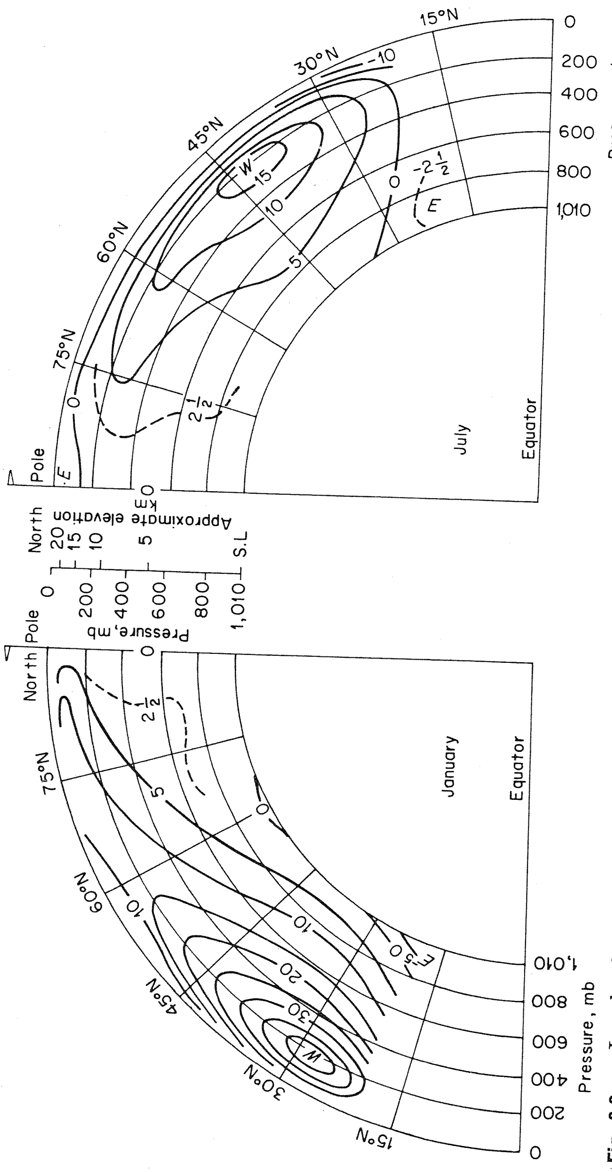


Fig. 8-8 Isovels of the zonal-average mean geostrophic wind (magnitudes in meters per second). (*From Y. Mintz and G. Dean, The Observed Mean Field of Motion of the Atmosphere, Geophys. Res. Paper*, 17, *Air Force Cambridge Research Center, Cambridge, Mass.*, 1952.)

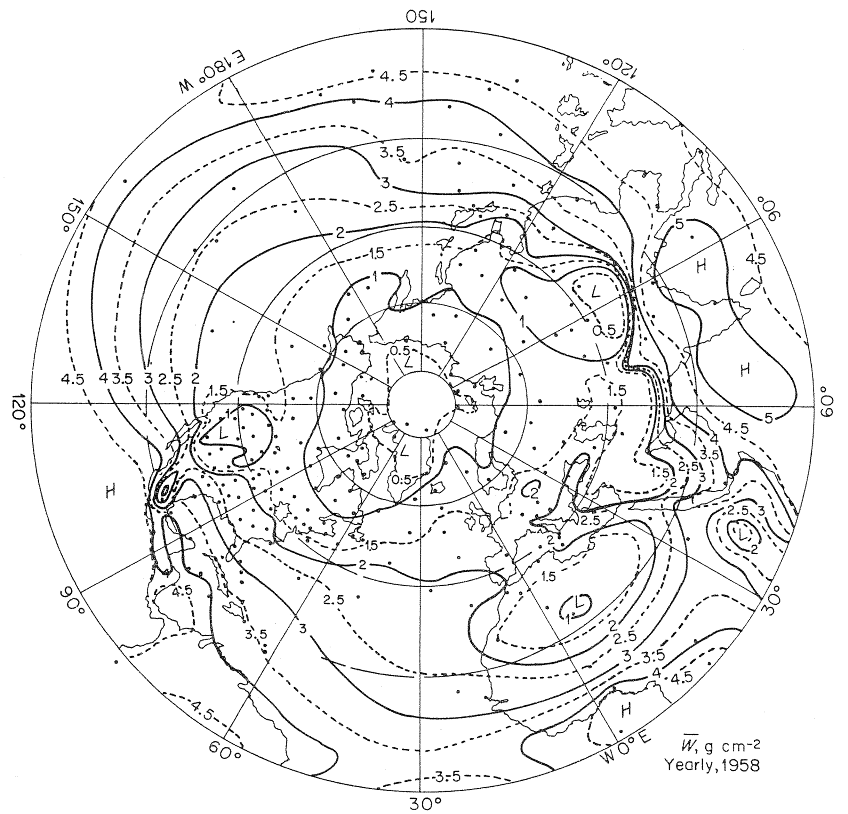


Fig. 8-9 Distribution of precipitable water in the atmosphere of the Northern Hemisphere for the year 1958. (*By permission from Ref.* 2.)

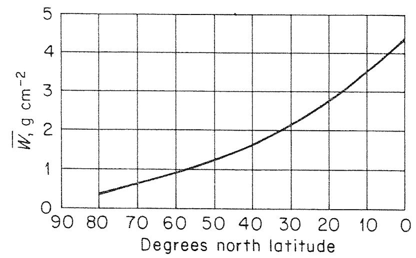


Fig. 8-10 Mean meridional distribution of zonally averaged precipitable water vapor. (*Data from V. P. Starr, J. P. Peixoto, and G. C. Livadas, On the Meridional Flux of Water Vapor in the Northern Hemisphere, Geofis. Pura Appl.-Milano*, vol. 39, 1958.)

ESR 320

1st Lecture
Section 4.3.1

2nd Lecture
Section
4.3.2 → End

4.3.1 GLOBAL AIR CIRCULATION

On the very longest scale, air circulation over the planet carries heat from the warm equatorial regions to the colder polar regions. Outside the atmosphere, the *solar constant,* or amount of power delivered by the Sun to a unit area oriented perpendicular to the Sun's rays, is approximately 1400 W/m^2 (based on Earth's mean distance from the Sun). Approximately half of this power is delivered directly to Earth's surface (see Fig. 2-30). This is an enormous rate of energy flow; averaged over the globe, it dwarfs the power output of all engines, generators, power plants, and furnaces combined. Solar heating of both the atmosphere and the oceans is greatest at low latitudes near the equator. The warmed fluids (both air and water) tend to rise due to buoyancy, and cooler fluid coming from the direction of the poles moves toward the equator to replace the warm, rising air or water. In exchange, the warmed fluid moves poleward. Planetary-scale convection cells thus are created in each hemisphere, in both the atmosphere and the oceans, and serve to transport heat toward the poles. If Earth did not rotate, one might expect relatively simple convective flow of fluids to occur, as shown in Fig. 4-11.

Because Earth does rotate, however, another set of forces also acts on the fluids; these *geostrophic* forces are the consequence of the fact that Earth's surface forms an accelerating reference frame. Relative to Earth's surface, all fluid motions are deflected perpendicular to their velocity by the *Coriolis force;* in the Northern Hemisphere, fluid motions are deflected to the right, and in the Southern Hemisphere, to the left. A parcel of fluid in motion in the Northern Hemisphere, under the influence of only the Coriolis force, experiences acceleration equal to

$$A = 2\,\omega(\sin\phi)v, \qquad [4\text{-}13]$$

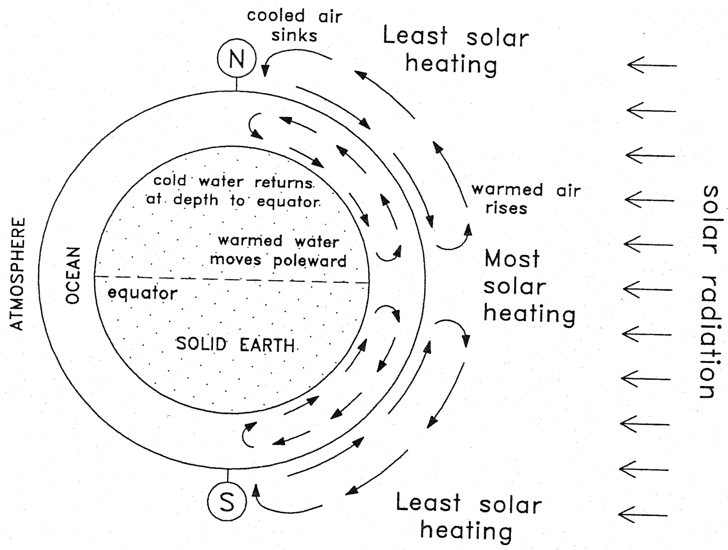
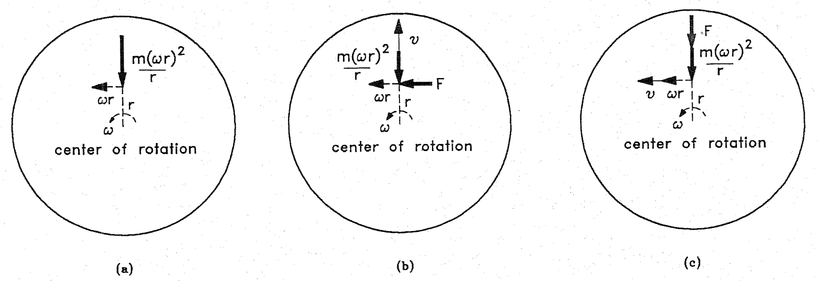


FIGURE 4-11 Global-scale tropospheric circulation as it would be if Earth did not rotate. Heat is transported from the equatorial area to the cold polar regions by both atmospheric and oceanic currents in each hemisphere.



(a)

- *No velocity v (relative to disk)
- *Absolute tangential velocity = ωr
- *Centripetal force = $(m(\omega r)^2)/r$
- *Centripetal force supplied by fluid

(b)

- *Radial velocity v (relative to disk)
- *KE = $1/2\ m(\omega r)^2$
- *$dKE/dt = m\omega^2 rv$ (n.b. $v = dr/dt$)
- *Power loss to centripetal force = $m\omega^2 rv$
- *Power input from F = $F\omega r$
- *Power input = power loss + dKE/dt
  $F\omega r = m\omega^2 rv + m\omega^2 rv$
- *$F/m = 2\omega v$

(c)

- *Tangential velocity v (relative to disk)
- *Absolute tangential velocity = $\omega r + v$
- *Total centripetal force = $m(\omega r + v)^2/r$
- *Neglecting small mv^2/r term,
  "extra" centripetal force = $F = 2m\omega v$
- *$F/m = 2\omega v$

FIGURE 4-12 Origin of the Coriolis force. Consider a small parcel of water of mass *m* within a body of fluid rotating on a disk. In (a), the parcel is not moving with respect to the disk, although its tangential velocity ωr is evident to a fixed observer, and a centripetal force of magnitude $m(\omega r)^2/r$ is being exerted by the surrounding fluid to keep the parcel on a circular path. The Coriolis force is the *additional* force that would have to be applied to the parcel to give it a uniform velocity v with respect to the rotating disk. Radial velocity and tangential velocity are considered separately in (b) and (c); any velocity can be expressed as a sum of these two components. Light lines represent velocity; heavy lines represent forces.

In the case of *radial* velocity (b), the magnitude of the required additional force can be calculated from the rate at which energy must be added to (or removed from) the parcel. (Recall that energy per time is power.) A fixed observer can see that the velocity of the parcel, and hence its kinetic energy, must increase as it moves radially outward and its tangential velocity increases.

In the case of *tangential* velocity relative to the disk (c), extra centripetal force (or less force, if the parcel is moving opposite to the direction of rotation of the disk) is required to keep the parcel traveling at a constant radius.

In each case, if the additional force is not applied to keep the parcel traveling in a straight path relative to the disk, the parcel will experience an acceleration of magnitude F/m with respect to the disk; this is the Coriolis acceleration. For this direction of disk rotation, the Coriolis acceleration is always to the *right* of the direction the parcel travels relative to the disk, and is always of magnitude $2\omega v$.

where A is the magnitude of acceleration [L/T^2], ω is the angular speed of Earth's rotation (2π radians/24 hr) [T^{-1}], ϕ is the latitude of the parcel of fluid, and v is the velocity of the fluid relative to Earth's surface [L/T].

The Coriolis force becomes particularly important when fluid motion occurs at large scales, as in vast lakes such as Lake Michigan (Fig. 2-6 b), or in atmospheric circulation (Fig. 4-12). Figure 4-12 explains the origin of the Coriolis force in terms of the radial and tangential components of the velocity of a fluid parcel in a rotating mass of fluid.

The Coriolis force, in conjunction with solar heating, creates a more complex global circulation pattern than that shown in Fig. 4-11. Three major latitudinal bands of surface winds result from these combined forces. In the Northern Hemisphere, the *trade winds* lie between the equator and approximately 30° latitude and are generally from the northeast (Fig. 4-13). North of the trade wind latitudes, between approximately 30° and 60° latitude, poleward-moving surface winds are deflected to the right by the Coriolis force, giving rise to the *westerlies*. Finally, from approximately 60° poleward, a third global-scale convective flow moves southward near Earth's surface and returns to the poles at higher altitude; the Coriolis force deflects these surface winds so that they come from the east (the *polar easterlies*), as seen by an earth-bound observer.

Prior to the 20th century, when trade across the Atlantic Ocean depended on sailing ships, ships leaving Europe often would sail south to pick up the

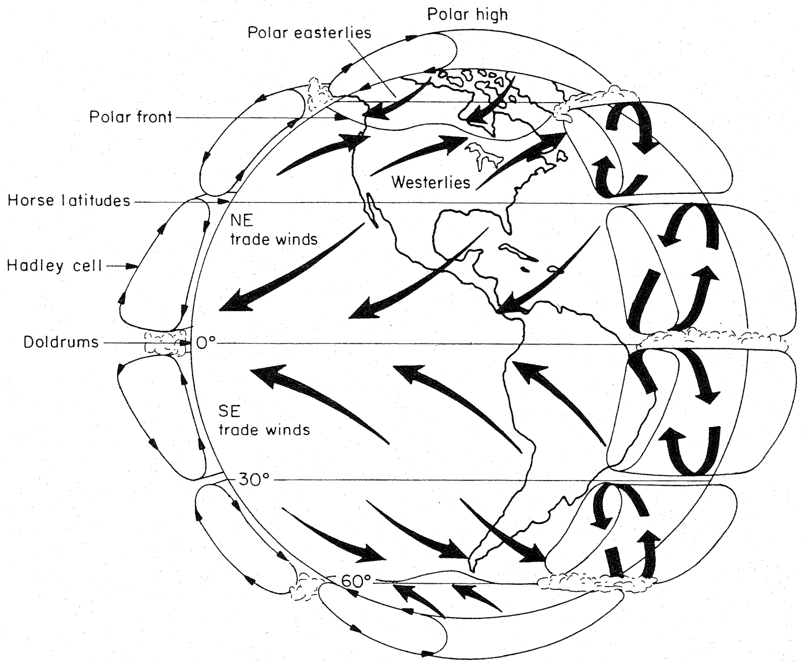


FIGURE 4-13 A simplified diagram of Earth's global-scale tropospheric circulation, including the Coriolis force. Note the three major bands of winds, as well as the cloud-producing effects of rising air near the equator and the polar fronts, and the desertifying effects of descending air at the poles and in the vicinity of 30° north and south latitude (adapted from Frederick K. Lutgens/Edward J. Tarbuck, *The Atmosphere: An Introduction to Meteorology*, 5th ed., © 1992, p. 170. Reprinted by permission of Prentice Hall, Englewood Cliffs, New Jersey.)

trade winds for the western crossing, and then sail north to pick up the westerlies for return to Europe (see Fig. 4-13). Following prevailing winds along lines of constant latitude was well suited to the large sailing ships of the time, with their square rigs optimized for downwind sailing. (It also was suited to early navigational tools that could accurately measure latitude, but could not determine longitude, due to the unavailability of sufficiently accurate timepieces (Sobel, 1995). Note that mariners used the conversion of 1 min latitude to 1 nautical mi. A *knot* is 1 nautical mph, or approximately 1.15 mph.)

These three cells of convective circulation occur within the troposphere, contributing to relatively rapid and complete mixing of this layer of the atmosphere. The tropical cells located north and south of the equator often

of the Hadley cells are often called *Ferrel cells*. Mixing occurs within the northern and southern hemispheres faster than between hemispheres, because there is no systematic large-scale, north–south advection of air in the vicinity of the equator, as shown in Fig. 4-13.

Many features of atmospheric chemical transport can be inferred by inspection of the global-scale atmospheric circulation pattern. The west-to-east movement of industrial chemicals, such as acid rain precursors, is a familiar example in the midlatitudes of the Northern Hemisphere. Of course, these belts of wind, such as the midlatitude westerlies, are subject to modification on a smaller scale by a variety of local and regional conditions. For example, a *wind rose* from Chicago (Fig. 4-14) shows that despite the dominant southwesterly direction of the wind, wind from other directions also occurs. To explain such departures from the average global circulation pattern, atmospheric processes on smaller scales must be considered.

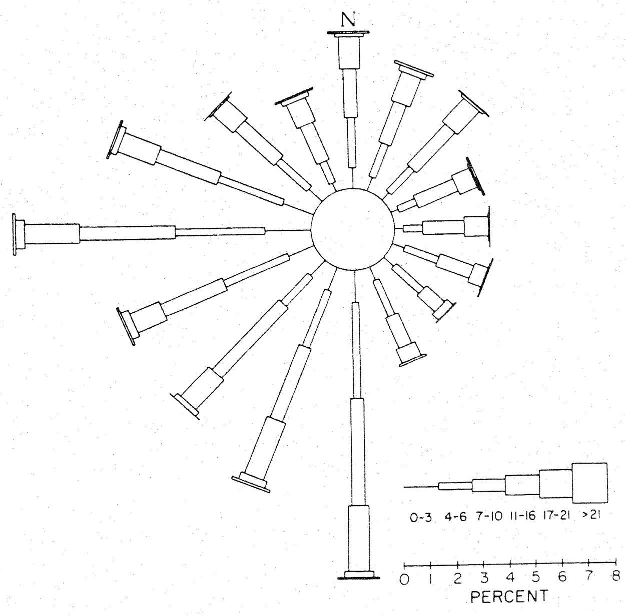


FIGURE 4-14 A wind rose, which graphically portrays the statistical pattern of wind velocities (in knots) at O'Hare Airport in Chicago from 1965 to 1969. Note the large fraction of time when winds are from the west to the south, as would be expected in the region of westerlies. At times,