# Part 9.  Tropical Cyclones

## 9.1  Introduction

The tropical cyclone is known as the hurricane in the north Atlantic and eastern north Pacific, typhoon in the western north Pacific, and just cyclone in the Indian ocean and Australia. One of the more colorful names it has been given is ‘The Serpent's Coil[1],’ a very appropriate name because from its coiled position the tropical cyclone strikes with a vengeance with winds in excess of 32 m/s (64 mi/hr), heavy rainfall, storm surges, and sporadic tornadoes. Born over warm, tropical, oceanic waters, the storm gains strength as it moves poleward and spins up into a cyclonically-rotating storm (counter-clockwise in the northern hemisphere) as it experiences the influence of the earth's rotation. At its maturity, the tropical cyclone is an inertially stable vortex whose circulations may extend to distances as much as 1600 km (890 mi) from the storm center. The highest sustained winds ever recorded in an Atlantic tropical cyclone occurred in Gilbert were 82.3 m/s (184 mi/hr) with a top gust measured at 199 mi/hr.  Not only do tropical cyclones produce winds that can rival a tornado in strength, but also they can produce extensive damage over large areas and for sustained periods.

## 9.2  Structure of the Mature Tropical Cyclone

A schematic of the three-dimensional structure of a mature tropical cyclone is shown in Figure 9.1.

[1] Farley Mowat, 1985:  *The Serpent's Coil*. Seal Books, McClelland and Stewart-Bantam Books, 222 pp.

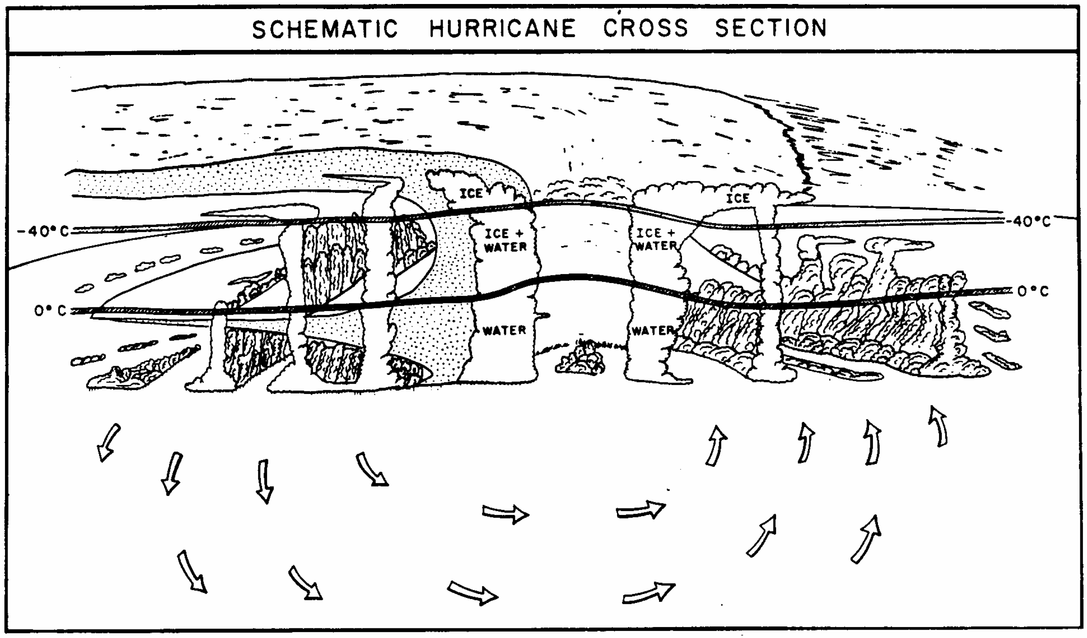



**Figure 9.1**. Schematic diagram of hurricane, showing low-level circulation and cloud types. The highest clouds, composed of cirrus and cirrostratus, occur at the tropopause, which is about 16 km. (From Stormfury, 1970: Project Stormfury Annual Report 1969, National Hurricane Research Laboratory, NOAA, AOML/Hurricane Research Division, Miami, FL, 20 pp.)

Near the surface, warm, moist air spirals inward toward the center of low pressure. At radii greater than about 400 to 600 km from the center, this flow is divergent, and sinking motion extends throughout most of the troposphere. This warm, sinking air is dry and usually is free of deep convective clouds, as seen in the satellite photograph of Hurricane Becky, which occurred in 1974 (Figure 9.2). Inside a radius of about 400 km, the low-level flow is convergent and the associated lifting of the warm, humid air produces extensive cumulonimbus clouds and precipitation. In spite of the relatively uniform tropical environment in which the tropical cyclone develops, it is characterized by a number of important mesoscale features. The mesoscale structure consists of the eyewall, a generally circular ring of intense convection surrounding the often cloud free eye; a region of stratiform cloud

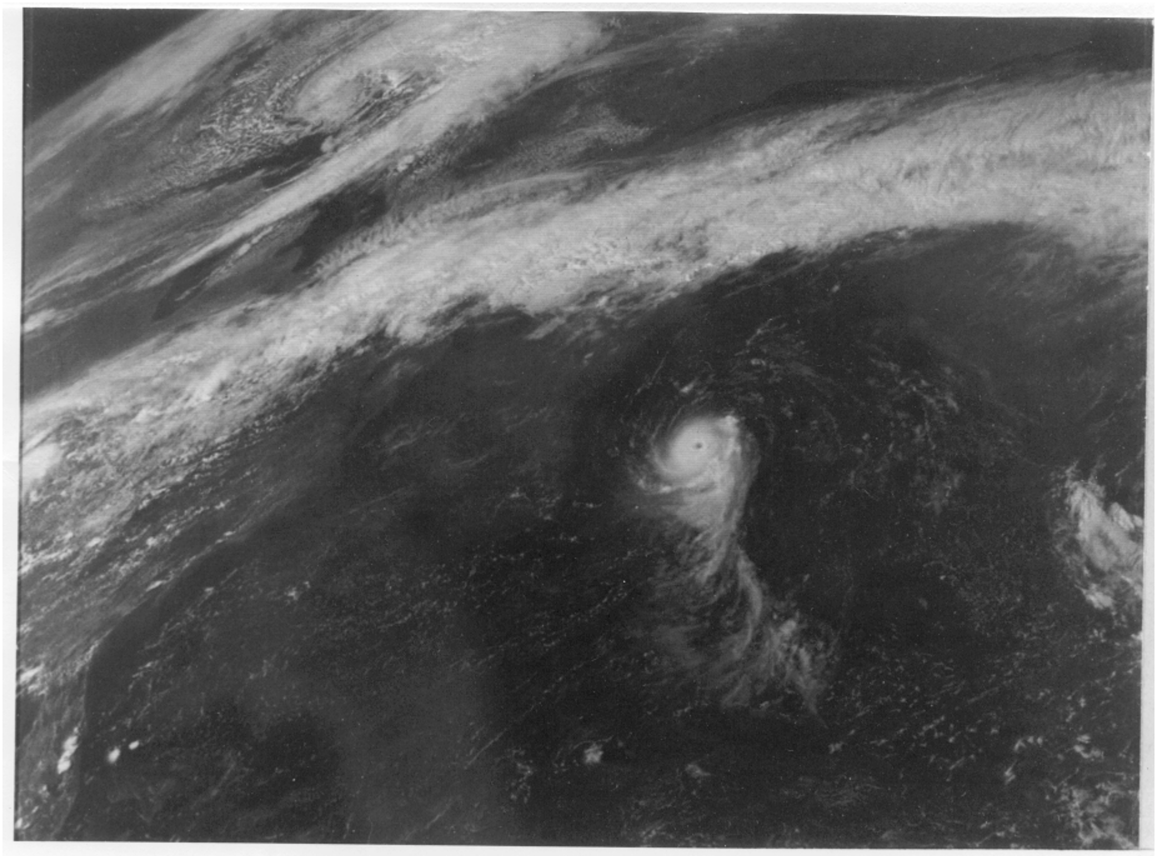
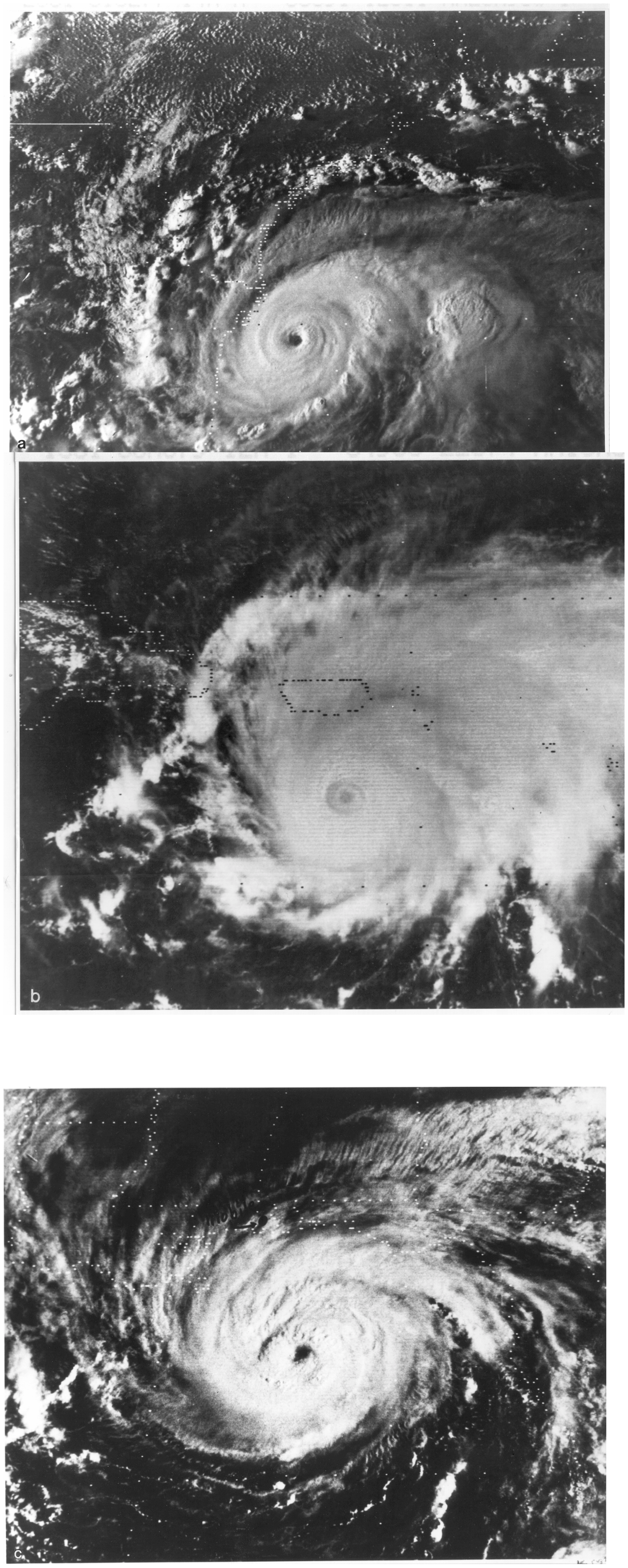


**Figure 9.2**. Satellite photograph of Hurricane Becky, 1800 UTC 20 August 1974. (Anthes, R. A., 1982: Tropical cyclones. Their evolution, structure, and effects. Meteor. Monograph No. 41, American Meteorological Society, Boston, 208 pp.)

and precipitation outside the eyewall; and spiral bands of convective clouds that assume various forms. Visible and infrared satellite imagery of four mature hurricanes shown in Figure 9.3 all show well-developed eyes and bands of clouds having variable structure.



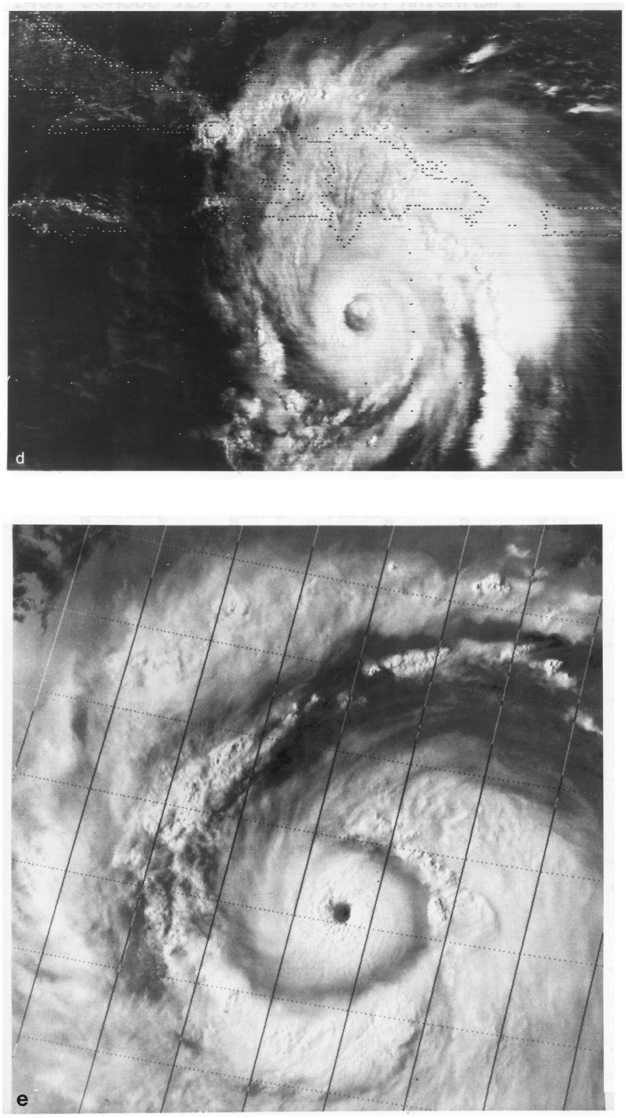



**Figure 9.3**. Satellite imagery of the four hurricanes near the time of flights by the research aircraft: (a) GOES-2 visible view of Hurricane Anita at 2300 UTC 1 September 1977 in the western Gulf of Mexico. (b) SMS-2 IR picture of Hurricane David at 1530 UTC 30 August 1979 south of Puerto Rico. (c) SMS-2 visible picture of Hurricane Frederic at 1300 UTC 12 September 1979 south of Mobile, Alabama. (d) SMS-2 IR view of Hurricane Allen at 1430 UTC 5 August 1980 south of Hispaniola.
(e) TIROS-N visible image of Hurricane Allen at 2115 UTC 8 August 1980 in the central Gulf of Mexico. (From Jorgensen, D. P., 1984a: Mesoscale and convective-scale characteristics of mature hurricanes. Part I. General observations by research aircraft. *J. Atmos. Sci*., **41**, 1268-1285.)

A schematic radar vertical cross section through the eyewall region is illustrated in Figure 9.4. The eye is depicted by generally sinking motions

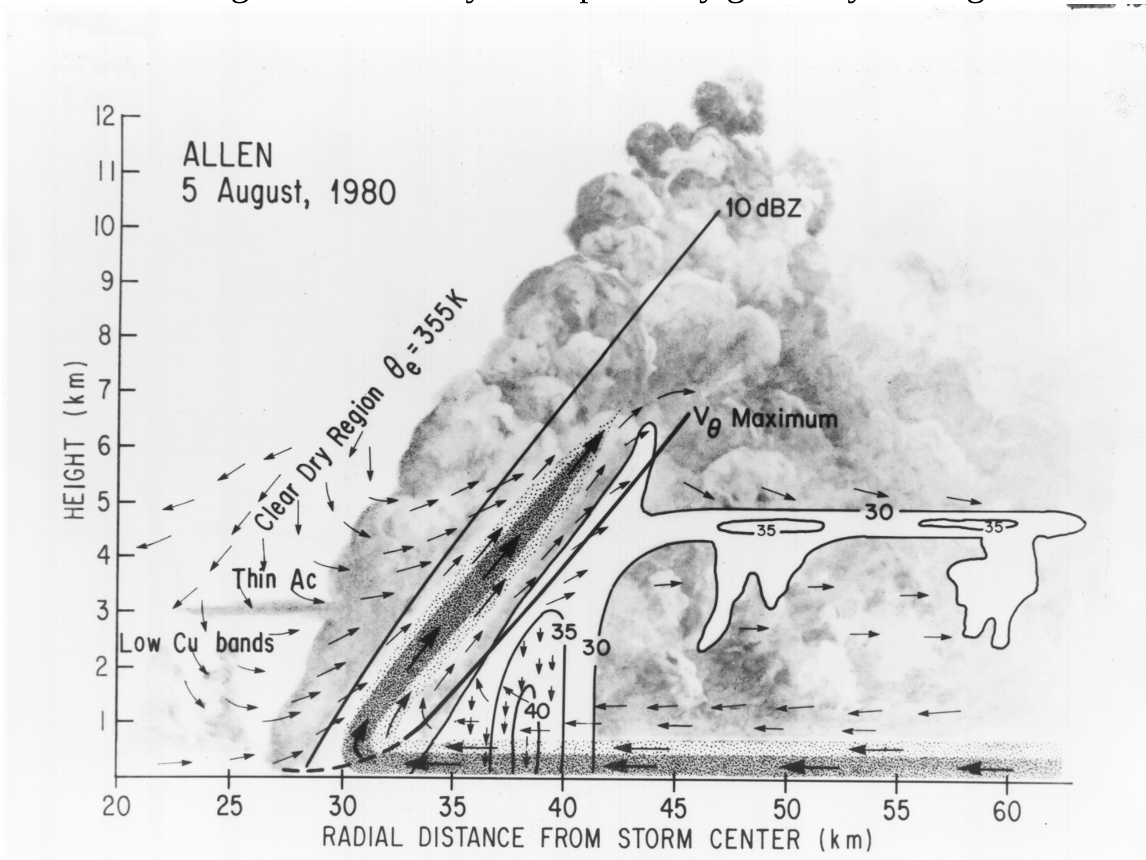



**Figure 9.4**. Schematic cross section depicting the locations of the clouds and precipitation, RMW, and radial-vertical airflow through the eyewall of Hurricane Allen on 5 August 1980. The slope of the cloudy region on the inside edge of the eyewall is based on radar minimum detectable signal analysis, aircraft altimeter readings, hand photography and observer notes. Darker shaded regions denote the location of the largest radial and vertical velocity. (From Jorgensen, D. P., 1984b: Mesoscale and convective-scale characteristics of mature hurricanes. Part II. Inner core structure of hurricane Allen. *J. Atmos. Sci.*, **41**, 1287-1311.)

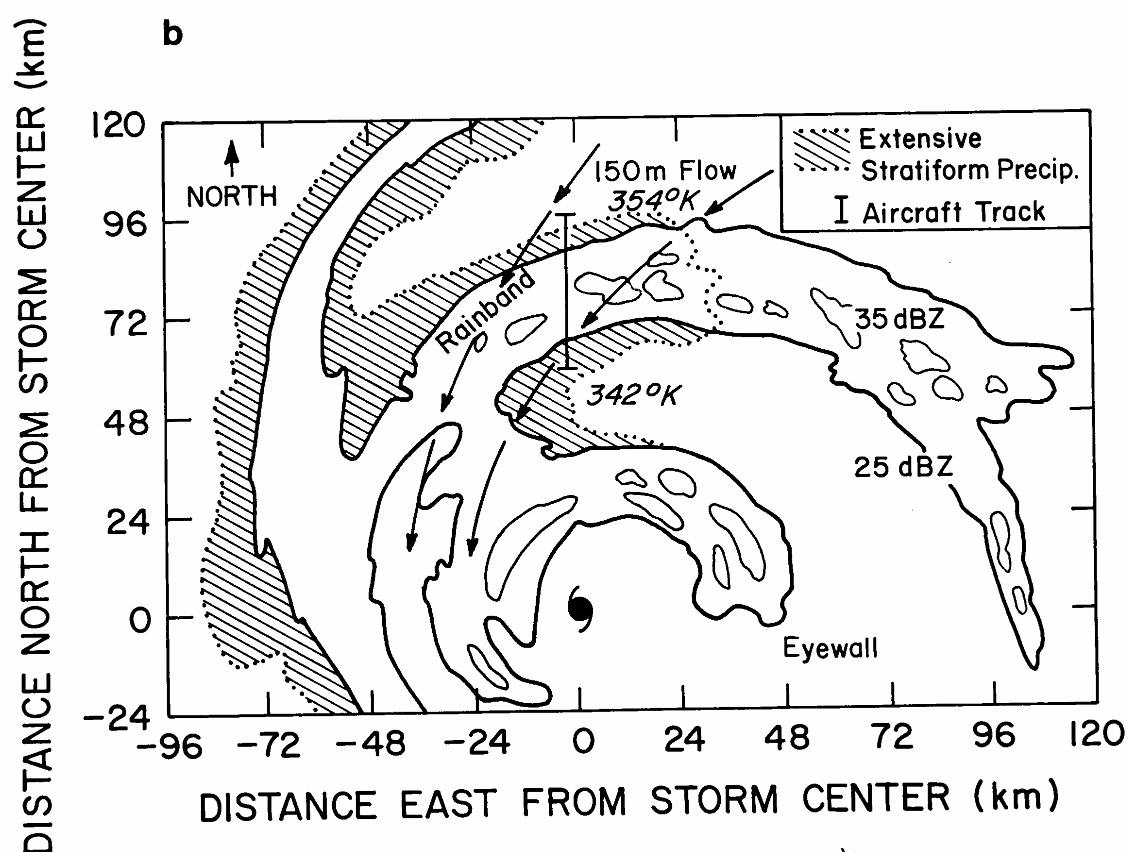



**Figure 9.5**. A horizontal radar depiction of rainbands. Low-level flow is also shown. (From Barnes, G. M., E. J. Zipser, D. Jorgensen and F. Marks, Jr., 1983: Mesoscale and convective structure of a hurricane rainband. *J. Atmos. Sci.*, **40**, 2125-2137.)

and is relatively cloud-free except for some low cumulus clouds and thin middle level cumulus clouds. Between 30 and 35 km outward from the center of the eye is the eyewall that is characterized by rapid ascent of the air rushing in toward the storm center. The eyewall tilts radially outward with height at an angle of about 30 degrees from the horizontal and, along with it, so also does the radius of maximum winds (RMW). The strongest radar reflectivity is located several kilometers outside the radius of maximum winds, while the maximum updrafts lie several kilometers inward of the RMW.

The region beyond the eyewall cloud is characterized by stratiform cloud and precipitation. The local regions of high radar reflectivity called *bright bands*, near 5 km height and outward from 45 km are due to melting of stratiform precipitation.

Beyond the eyewall region, the predominant mesoscale features of the tropical cyclone are spiral bands of clouds and rainfall. The rainbands are typically 5 to 50 km wide and spiral inward toward the center over radial distances of 100 to 300 km. Cumulus clouds tend to form on the inside of the bands and move cyclonically (counter-clockwise in the northern hemisphere) around the storm center. The convective elements typically move outward from the storm's central regions during the formative stage of the storm and dissipate further away from the storm center. The bands can be classified as two types: propagating, wavelike bands associated with internal gravity waves and groups of bands that remain in a nearly fixed location relative to the storm center. Figure 9.5 illustrates two rainbands as seen by radar outward from the eyewall.

Like a giant flywheel, the region within 100 km of the storm center is inertially stable and is not affected strongly by outside weather systems. At distances of 100 km to more than 1000 km from the storm center, the flow is not so inertially stable and larger-scale weather systems can easily affect the airflow in these regions. At these distances from the storm center, the flow is largely in gradient wind balance but approaches geostrophic balance at the outer cyclone limits. The dimensions of tropical cyclones mentioned above are fairly typical. It must be recognized, however, there is enormous variability in the size of tropical cyclones with some smaller storms being able to fit within the eye of the largest storm.

## 9.3 Energetics of the Mature Storm

The primary energy driving the tropical cyclone comes from the sea. Air flowing over a warm ocean surface receives energy from the surface primarily in two forms: sensible heat and latent heat. The transfer of sensible heat from the ocean surface refers to the direct transfer of heat from the warm ocean surface to cooler overlying air by small-scale turbulent eddies. The rate of transfer of heat from the ocean surface is greater the larger the temperature difference between the ocean surface and the overlying air and the stronger the wind speed near the ocean surface.

The transfer of latent heat from the ocean surface refers to the moistening of the air near the ocean surface by evaporation of seawater. In general, the rate of latent heat transfer by evaporation from the ocean surface increases as the temperature of the ocean surface is warmed and, as with sensible heat, with the wind speed. The energy required to evaporate seawater comes from the ocean itself, which is a large heat reservoir. Thus, moisture is supplied to the air immediately over the ocean surface without any substantial cooling of the air.

As shown in Figure 9.6, as low-level air is swept inward towards the center of the cyclone, it collects sensible and latent heat from the ocean surface at increasingly large rates because the wind speed increases toward the eyewall region.

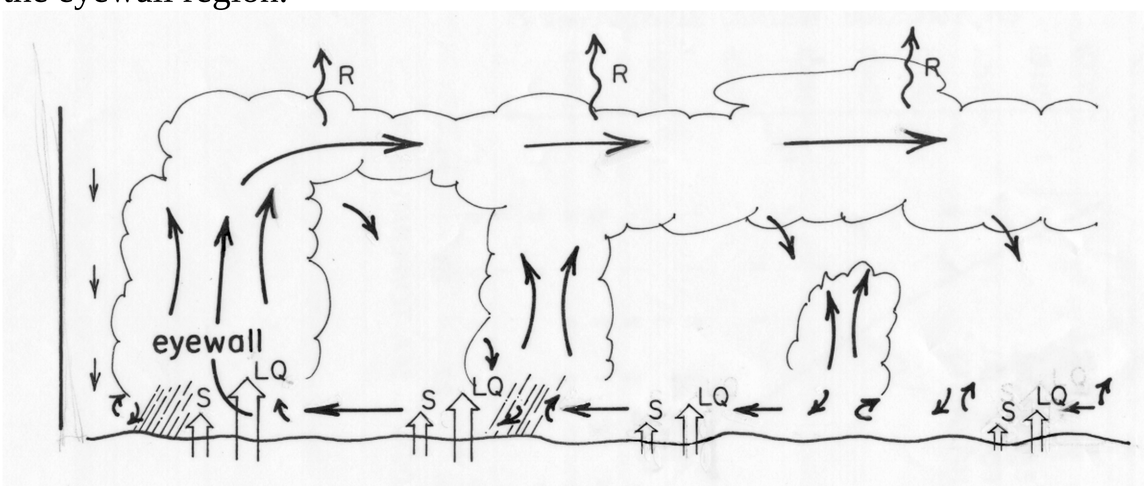



**Figure 9.6**. Illustration of sensible (S) and latent heat fluxes (LQ) from the ocean surface beneath a tropical cyclone. As winds increase in speed toward the central region of the cyclone, fluxes of sensible and latent heat increase substantially. Cumulus clouds and sub-cloud eddies transfer the heat supplied by the ocean surface upwards to the middle and upper troposphere. In the upper troposphere and lower stratosphere heat is lost to space by longwave radiation emissions (R) from the tops of the stratiform-anvil clouds.}

Small-scale turbulent eddies transfer the heat and moisture upwards to levels where it becomes saturated and cumulus clouds form. By condensing water to form cloud droplets and as a consequence releasing latent heat, the moisture or latent heat transferred from the ocean surface warms the cloudy air at a rate, which is roughly proportional to the precipitation rate in the clouds. Thus, cumulus clouds and sub-cloud eddies transfer the sensible and latent heat from the ocean surface to the middle and upper troposphere. As the air moves outward from the central regions of the storm in the stratiform-anvil clouds, much of the energy gained at the ocean surface is radiated to space by infrared radiation. Emanuel[2] has likened the tropical cyclone to an idealized heat engine, called a *Carnot engine*, which is very efficient in converting heat into work or kinetic energy of motion (in this case winds). In a Carnot engine, heat is input at a single high temperature and all the heat output is ejected at a single low temperature. The amount of work produced by the Carnot engine is proportional to the difference between the input and output temperatures, and is the maximum amount of energy that can be extracted from a heat source. In the case of a tropical cyclone, the amount of work or strength of the winds in a storm is proportional to the difference in temperature between the heat input level or the ocean surface and the heat output level, or the tops of the anvil clouds. In the case of the Carnot engine, heat input is in the form of sensible heat, while in a tropical cyclone, heat input is in the form of sensible and latent heat, with latent heat being the dominant contributor to the storm energetics. The tropical cyclone can thus be viewed as a heat engine, although less efficient than the idealized Carnot engine. Figure 9.7 shows that tropical cyclones do indeed increase in strength as sea surface temperature increases.

[2] Emanuel, Kerry A., 1988: Toward a general theory of hurricanes. *American Scientist*, **76**, 371-379.

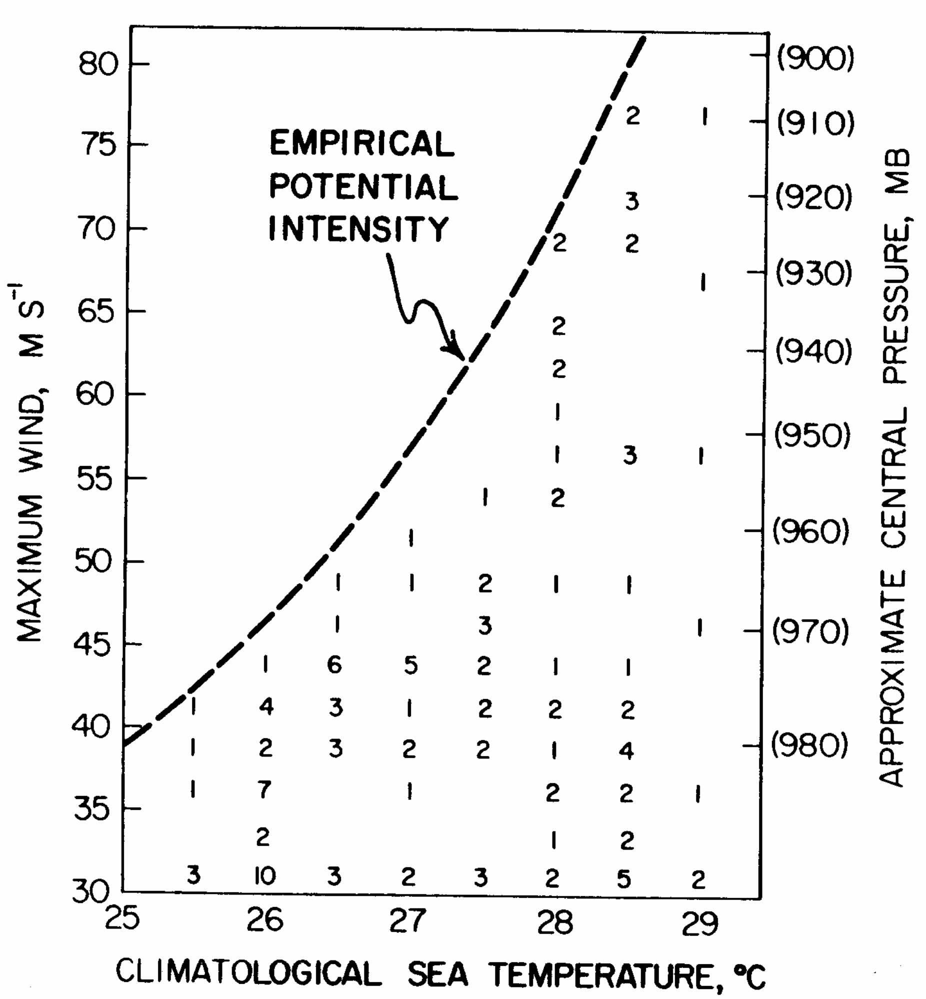



**Figure 9.7**. Empirical relationship between climatological sea surface temperature (SST) and best track for a sample of hurricanes. Small numbers are the frequencies of occurrence of a given intensity and SST, and the dashed line is the empirically determined upper bound on hurricane intensity. (From Merrill, R.T., 1988: Environmental influences on hurricane intensification. *J. Atmos. Sci.*, **45**, 1678-1687.)

Another energy source for a tropical cyclone is a result of the cooperative interaction between deep convective cells and the cyclone circulation. We have seen from preceeding chapters that low-level convergence of warm, moist air is necessary for maintaining cumulonimbus cells. We have also seen that above the boundary layer, flow about a low pressure center can approach gradient wind balance, in which the horizontal pressure gradient force is balanced by the sum of the Coriolis and centrifugal forces. Within the boundary layer, friction slows down the winds and the flow about a low pressure center adjusts to a new balance of forces in which the flow turns toward low pressure. The turning of the winds toward low pressure in the boundary layer causes convergence called Ekman convergence.

In a tropical cyclone, the frictionally-induced convergence can supply cumulus clouds with warm, moist air. The warm, moist air could either be resident at low levels in the cyclone environment or it could be transferred from the ocean surface as described above. Fed by the supply of warm moist air, deep, precipitating convective clouds develop, which release latent heat and warm the air in the cloudy regions. The column of warmed air lowers the pressure at low-levels in the storm, which in turn favors the development of increased boundary layer convergence. This, then fires up more intense convection, and so on. This process is referred to as conditional instability of the second kind (CISK). (Conditional instability of the first kind refers to free parcel.) A schematic illustrating the development of a tropical cyclone from an environment characterized by a weak, low-level cyclone and upper-level anticyclone is shown in Figure 9.8.

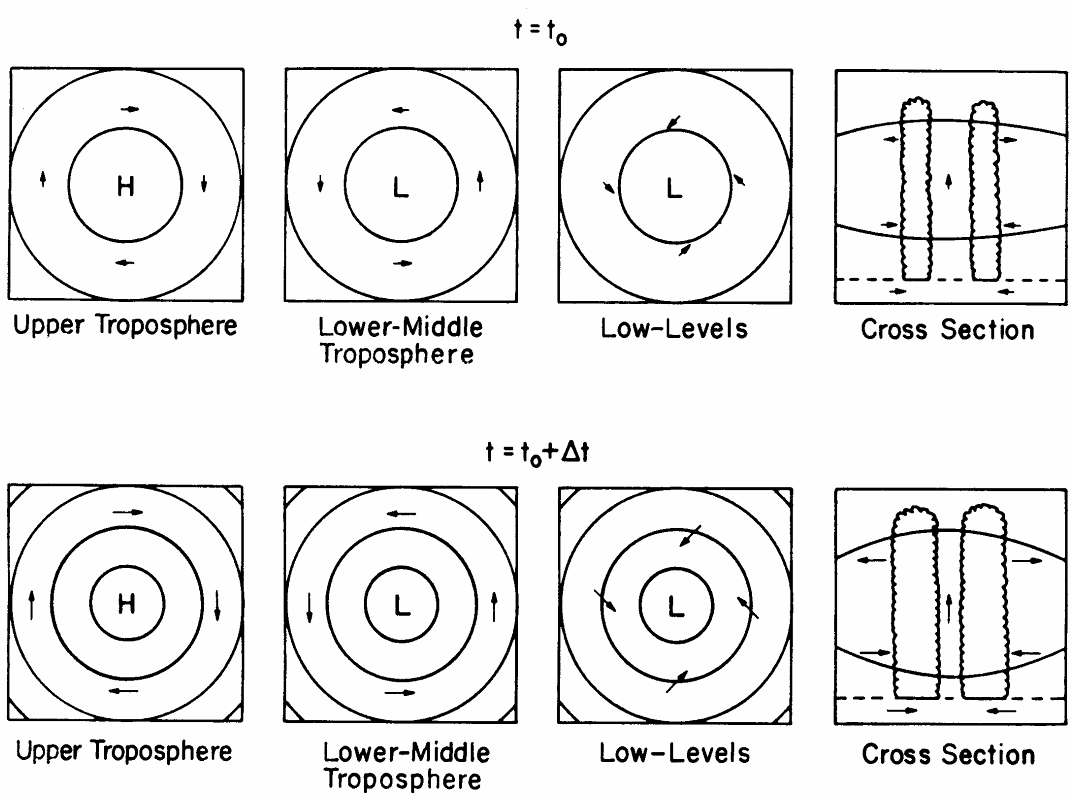



**Figure 9.8**. Schematic depiction of CISK. Initially there is an anticyclone aloft and a cyclone in the lower troposphere. Ekman convergence in the

boundary layer provides moisture to a population of precipitating clouds. Most of the diabatic heat source is balanced by the adiabatic cooling associated with the induced transverse circulation. However, there is a slight warming, yielding a more intense anticyclone aloft, a more intense cyclone below, increased Ekman convergence, more clouds, etc. (Adapted from Schubert, W.H. and J.J. Hack, 1982: Inertial stability and tropical cyclone development. *J. Atmos. Sci*., **39**, 1687-1697.)

Scientists are still debating whether a tropical cyclone is driven primarily by a CISK-type process or by a Carnot engine-type process or perhaps some other processes. There is strong evidence that the strength of tropical cyclones and the intensity of the surface low pressure cannot be achieved by a CISK mechanism alone. Boundary layer fluxes of heat and moisture are necessary for maintenance and perhaps intensification of an incipient cyclonic storm to hurricane strength. Moreover the presence of large numbers of cumulonimbus clouds in the tropics drives the tropical cyclone environment to near wet adiabatic in structure and as a result there is little conditional instability to drive a strong CISK mechanism. On the other hand, large regions of the tropical ocean exhibit sea surface temperatures high enough to sustain a tropical cyclone by a Carnot cycle-type process alone, yet few, if any, tropical cyclones form. This means that some other factors must come into place to create a storm of hurricane intensity.

## 9.4 Genesis of Tropical Cyclones

There are favored regions on the earth for tropical cyclone genesis. Figure 9.9 illustrates they are in the western Atlantic, eastern Pacific,

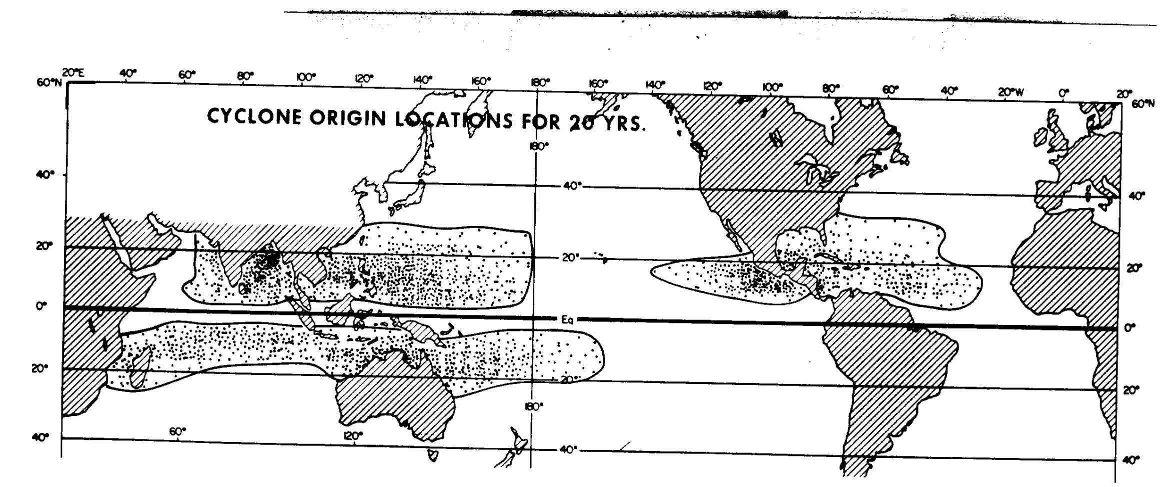



**Figure 9.9**. Locations of the regions of genesis of tropical cyclones for a 20-year period. (Gray, W.H., 1975: Tropical cyclone genesis. Dept. of Atmos. Sci.Paper No. 234, Colorado State University, Dept. of Atmospheric Science, Fort Collins, CO 80523.)

western North Pacific, North Indian Ocean, South Indian Ocean, in northern Australia and the South Pacific. Notable is the complete absence of tropical cyclones in the South Atlantic and eastern South Pacific, presumably due to the cooler ocean temperatures in those regions. The importance of the earth's rotation is evidenced by the fact that few tropical cyclones form equatorward of 4-5 degrees latitude. About 65% of all cyclones form between 10\deg and 20° of the equator.

Most tropical cyclones form in late summer and early autumn, when ocean temperatures are at their warmest. An exception is the western north Pacific, where cyclones have been observed in every month. Sea surface temperature is obviously an important factor in tropical cyclone genesis as few storms are observed where the sea surface temperatures are less than 26.5°C (80°F). Figure 9.7 suggests that some weak, tropical cyclones do form at slightly cooler temperatures, however.

In a particular region, tropical cyclones do not form uniformly throughout the cyclone season, but instead, tend to form in episodes of increased activity. Cyclones tend to form in 2- to 3-week active periods followed by 2- to 3-week inactive periods. It is generally believed that the long term active and inactive periods correspond to the well-known 40 to 50 day oscillation in tropical pressures, upper-level winds, and cloudiness.

Normally, tropical cyclogenesis occurs in a region in which a pre-existing mesoscale disturbance exists. This disturbance may be characterized by weak, low-level convergence and cyclonic vorticity. The low-level convergence fuels convective clouds with warm, moist air, which favors the development of convection, organized on the mesoscale. Furthermore, if the weak region of convergence intensifies, then in a conditionally unstable atmosphere, the pre-existing vorticity can intensify through the convergence mechanism.

As already mentioned, the atmosphere in which tropical cyclones develop is not noted for large values of conditional instability. Weak conditional instability is present, which will support thunderstorms of weak intensity but nothing like the instability that exists over the central United States when there are severe weather outbreaks. Of course, the instability over the central United States strengthens and decays over a diurnal cycle and cannot, therefore, sustain cyclogenesis for periods of many days. A deep, moist layer in the lower troposphere is also present when tropical cyclone genesis occurs. Both these conditions favor the development of ordinary cumulus clouds and weak cumulonimbus clouds rather than intense, rapidly propagating squall lines. Likewise, the environment of developing tropical cyclones is characterized by weak, vertical shear of the horizontal wind. Again, weak vertical shear favors the formation of ordinary cumulus clouds and thunderstorms rather than rapidly propagating squall lines. This is perhaps due to the fact that the more rapidly propagating modes of convection will move quickly away from regions of low-level support, such as convergence. As a result, latent heat released in these storms mainly generates gravity waves that propagate away from the incipient cyclone. In weak shear, the ordinary thunderstorms will remain coupled with the region of low-level convergence and thereby contribute to the heating of the embryonic disturbance and to the lowering of pressure in the storm. This lowered pressure will contribute to strengthened convergence and further development of the cyclone.

The final factor that appears to be important to the genesis of tropical cyclones is the presence of divergent flow at upper levels. Upper-level divergence essentially removes air mass from an air column, thus making it easier for surface pressure to fall, and for low-level convergence to form, (see Fig. 9.10a).

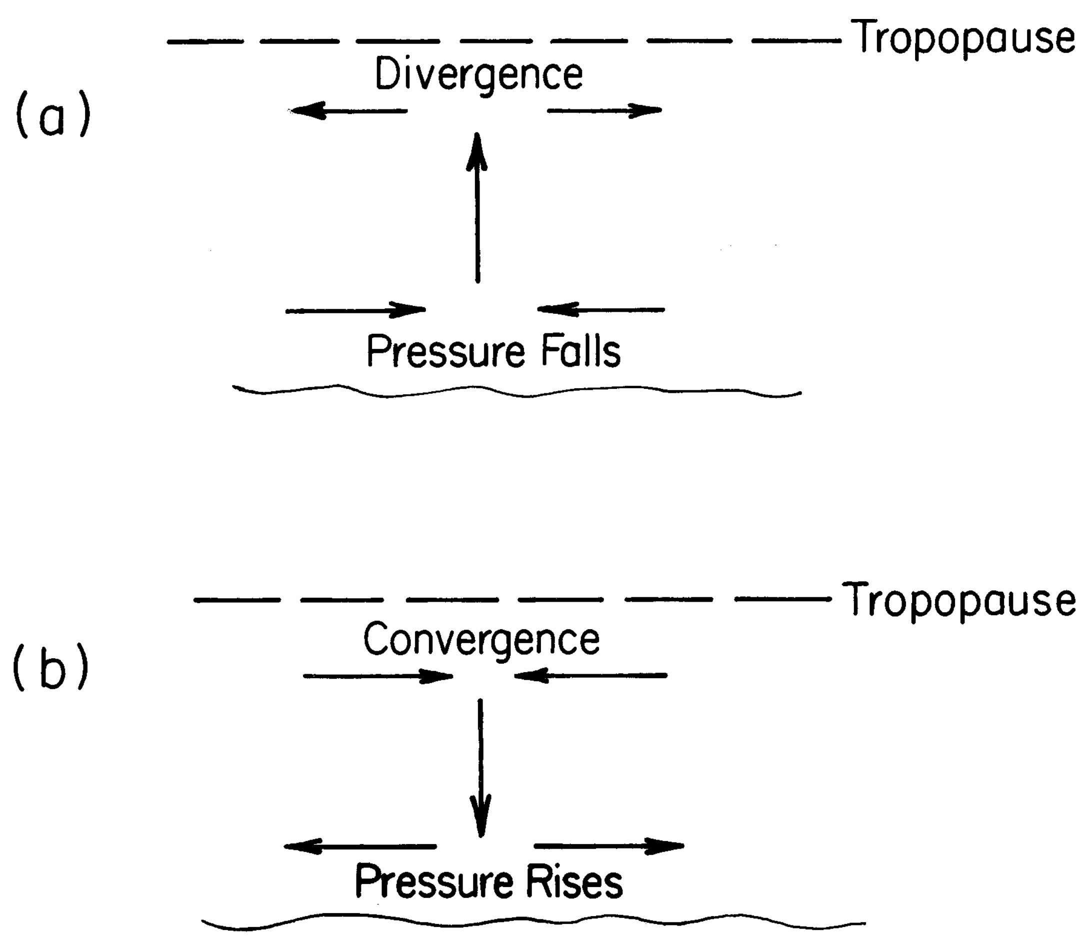



**Figure 9.10**. Schematic of the relationship between upper level divergence (a) or convergence (b) and variations in surface pressure and vertical motion in the troposphere.

The development of cumulus convection is aided when upward motion forms, which further favors lowering of surface pressure, increased convergence, and spin-up of the system. Conversely, as shown in Figure 9.10b, upper-level convergence causes surface pressure to rise and air to weakly descend through the troposphere. The descending air causes adiabatic warming, stabilizing the atmosphere and inhibiting the formation of cumulus clouds.

In summary, the conditions favorable for tropical cyclogenesis are as follows:

- A location poleward of about 4 to 5\deg latitude.
- Warm sea-surface temperatures.
- A pre-existing atmospheric disturbance with upper tropospheric divergence, low-level convergence, and cyclonic vorticity.
- Weak conditional instability and a deep layer of moist air.
- Weak vertical shear of the horizontal wind.

In contrast to a mesoscale convective complex, which may form in a time period of 6 to 8 hours, a tropical cyclone may require 5 or 6 days to reach hurricane intensity. Let us consider the hypothetical genesis of a tropical cyclone over the north Atlantic.

The first stage of a typical tropical cyclone genesis occurs when a mesoscale storm forms over the strongly heated arid regions of the African continent. The storm is characterized by a shallow layer of low-level convergence and weak, cyclonic vorticity at low levels. Under the influence of the prevailing easterly flow, the disturbance moves off the west African coast over the sea, where water temperatures are relatively cool, below the threshold temperature of 26.5°C, where cyclogenesis is favored. In the upper troposphere, weak sinking motion tends to suppress convection.

For the next several days, the disturbance drifts slowly westward with little indication of strengthening. Convective showers form, but heating in the convective showers is largely compensated by adiabatic cooling in the slowly rising, low-level air. The system remains cool in the middle and upper troposphere, and surface pressures change little.

On the third day over sea, the disturbance moves over warmer sea surface temperatures. The enhanced flow of moisture from the warmer sea fuels deep, convective clouds, which release latent heat, lowering surface pressures, causing increased low-level convergence, increased low-level vorticity, and slow, rising motion in the lower troposphere. The slow, rising motion and convective clouds moisten the lower troposphere. Because the disturbance remains beneath a region of weak, descending air in the upper troposphere and moderately strong shear, the disturbance does not strengthen appreciably for the next several days.

In its slow, westward path, the disturbance encounters sustained warm sea-surface temperatures, and finally, on the sixth day, the disturbance moves under a region of favorable upper tropospheric divergence and weak vertical shear of the horizontal wind. A chain of events then ensues, in which upward motion intensifies in the middle troposphere. The convective cells now find themselves in a favorable environment and increase in activity. Heating from the precipitating convective clouds lowers the pressure at low-levels, which increases low-level convergence.
The strengthened low-level convergence in a weak, cyclonic vorticity field increases low-level vorticity by the convergence mechanism. Strengthened winds rushing into the center of low pressure pick up moisture and heat from the sea surface at greatly increased rates. The convective clouds transport the warm, moist air aloft, where the latent heat released warms the air column. At upper levels, the warm moist air forms a larger stratiform anvil cloud, where pronounced longwave radiative cooling occurs at the top of the cloud layer. The disturbance now evolves into a Carnot engine-like machine with heat input at warm temperatures at low-levels, and heat output at cold temperatures at upper levels. A rapidly spinning tropical cyclone has been born!

## 9.5 Tropical Cyclone Motion

Once a mature tropical cyclone has been identified, the forecaster must now predict the motion of the storm. Unfortunately, tropical storm motion is a result of many complicated interactions among the external flow fields and the variations in the internal dynamics of the storm. Figure 9.11 illustrates the tracks of north Atlantic hurricanes in August for a 98-year period. The figure shows that most hurricanes are spawned off the African coast between

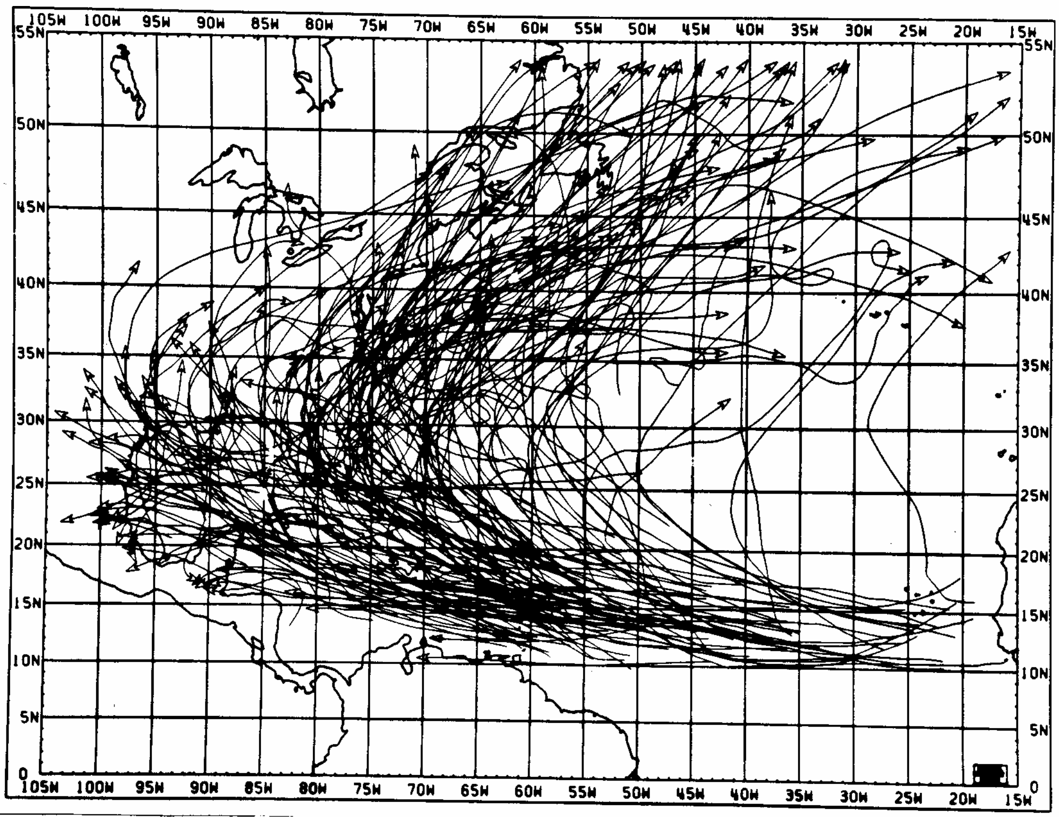



**Figure 9.11**. Tracks of North Atlantic hurricanes or tropical storms during 1886-1984, beginning in August (195 storms). (Courtesy of C. Neumann.)

10 and 17° north latitude and then make a generally westward track, turning northward between 60 and 85° west longitude. A closer examination of the figure shows there is considerable variability in the tracks with some storms continuing westward into the Mexican coast while others take abrupt northward tracks, missing landfall altogether or ramming into the south Texas coast. Some tracks follow a consistent, slightly curving path, while others make abrupt turns and even reversal in direction.

A common approach to hurricane track forecasting is to assume that the storm moves like a solid spinning cork in a stream. The forecaster must identify a ‘steering’ level in the atmosphere, which pushes the cyclone through the atmosphere. The undisturbed flow at middle tropospheric levels is generally a good steering level. It appears that the steering concept can account for 70 to 80% of the variability of storm motion in the Atlantic and Australian regions. Nonetheless operational determination of steering currents is complicated by the lack of data over oceanic regions and because larger scale pressure patterns change in time, so that the orientation and strength of the steering currents change in often-unpredictable ways. Furthermore, when the steering currents are weak and poorly defined, tropical cyclone tracks can be quite erratic in behavior.

The generally westward movement of tropical cyclones occurs even in the presence of weak or nonexistent easterly flow at middle levels. This is because of the variation of angular velocity across the north/south extent of the cyclone. We have seen that the effect of the earth's rotation is greater at higher latitudes than at lower latitudes. As a result, in the northern hemisphere, air flowing from north to south on the western side of a cyclone carries strong, earth-angular momentum southward; turning the southward flow to the right or to the west. On the storm's eastern flank, air moving northward carries little earth-angular momentum and, as a result, creates only weak turning of the winds to the right into an easterly direction. The net result is a generally westward drift of the tropical cyclone even when steering currents are weak.

The mass and momentum transport by thunderstorm cells within the tropical cyclone can also affect the structure of the steering flow. The mass that is transported aloft by thunderstorms flows outward in the upper troposphere and descends at some distance away from the cyclone. Generally, the mass outflow is not distributed uniformly in the upper troposphere, but is instead concentrated in localized regions called *outflow jets*. If, for example, the outflow jet concentrated mass ahead of the moving storm, a pressure high would build up, slowing the steering-current winds and the movement of the storm. A concentration of mass on the right, forward flank of the storm can frequently result in a sharp turning of the storm to the left of its expected track.

Upon encountering land masses, especially with mountainous terrain, the cyclone can exhibit unpredictable changes in movement from those expected from steering-level flow concepts. As airflow circulating about a cyclone encounters complex terrain, the flow may be blocked, causing local accumulations of mass on the windward side of the mountains. The altered pressure fields can then change the direction and strength of the steering flow, again resulting in abrupt changes in storm motion.

## 9.6 Decay of the Tropical Cyclone

The decay of a tropical cyclone generally coincides with landfall or the northward penetration of the storm into middle latitudes. As the storm moves over land, it loses its supply of warm, moist air, which we have seen is important to sustaining the cyclone as an efficient machine. Moreover, the topography and general roughness of the surface due to vegetation increases surface friction, which decreases the winds in the storm. Movement over smaller land masses may result in only weakening of the storm strength, while penetration inland of major continents can lead to the demise of the cyclone.

As shown in Figure 9.11 many tropical cyclones turn northeastward as they approach the North American continent. Often the northeastward track carries the storm over the Gulf Stream, which sustains a supply of warm, moist air into the cyclone. Eventually, the northward drift of the storms brings them in contact with cooler ocean temperatures and increases the likelihood that they may encounter polar airmasses. The interaction with cold fronts and the strong vertical shear of the horizontal wind associated with polar fronts can alter the structure of a tropical cyclone, transforming it from a nearly symmetric circulation to a more asymmetric one characteristic of extra-tropical cyclones. Typically, such an encounter leads to cooling of the storm core and a rise in surface pressures. Occasionally, the storm may intensify for a short time but eventually the loss of the tropical cyclone characteristics will prevail and the storm will weaken and be no longer recognizable as a tropical cyclone.

## 9.7 Tropical Cyclone Damage

The tropical cyclone is most notorious for the strong winds it produces. Wind damage is roughly proportional to the square of the wind speed. Therefore, modest increases in wind speed can result in a major enhancement of storm damage. In the northern hemisphere, wind damage is the greatest in the right, forward quadrant of the storm with respect to its direction of motion. It is in this quadrant that the cyclonic or counter-clockwise rotating flow about the storm center combines with the motion of the storm to produce the strongest wind speeds. Wind damage can be greatly enhanced in local regions by what are called wind squalls, where local winds increase in speed substantially above normal or average wind speeds. Such wind squalls can occur when convection in the eye wall region undergoes explosive development. In some tropical cyclones, a jet of warm, dry, middle-level air is observed to enter the cyclone, which, when transported over the low-level warm, moist air, produces locally unstable soundings not too different from soundings associated with severe thunderstorms over the central United States. The resultant convective bursts are so intense that the updrafts produce cloud turrets, which penetrate deeply into the stably stratified stratosphere and eject anvil clouds that completely obscure the eye of the cyclone. In some storms, such convective bursts have been observed to occur repeatedly for as long as 12 to 24 hours. Surface wind speeds associated with the convective bursts may be twice as large as wind speeds associated with the normal cyclone circulation. Local wind squalls associated with hurricanes are also found in the convective regions along the main rainbands of the cyclone and in squall lines in the pre-cyclone environment.

We noted in Chapter 4 that hurricanes are frequently the progenitors of tornadoes. It is believed that strong, low-level wind shears favor the development of tornadoes by the tilting mechanism. Tornadoes appear to be the most common, in the northern hemisphere, in the right front quadrant of storms that are about to make landfall. Tornadoes normally develop in the outer edge of the eye wall region or in the outer rainbands of the storm.

The sustained hurricane-force winds can also produce monstrous ocean waves. Winds of hurricane strength (32 m/s) can be expected to produce wave heights in excess of 19 m (62 ft), causing havoc to shipping and major damage along shorelines. As the cyclone approaches the shoreline, the strong winds pile up the ocean water along the coast. The decreased depth of the ocean and the low pressure about the center of the storm all contribute to a *storm surge*, especially on the right forward quadrant of the storm (in the northern hemisphere), where wind speeds are strongest. The combination of strong, sustained winds associated with major hurricanes and local coastal features, such as bays, can lead to elevation of sea levels in excess of 10 m (30 ft) and major flooding.

The torrential rains in tropical cyclones can also lead to extensive flooding. In some areas, however, the rainfall from tropical cyclones is a major contributor to seasonal rainfall for the survival of livestock, wildlife, plants, and rivers and reservoirs for human use. Unfortunately, we cannot order the optimum amount of rainfall, and some storms have been known to dump in excess of 48 cm (19 in) over a several day period. Even whe n a tropical cyclone has been downgraded to a tropical disturbance category, the penetration of the remaining circulation and precipitation bands into mountainous regions can lead to extensive flooding.