16 TROPICAL CYCLONES

Contents

Intense synoptic-scale cyclones in the tropics are called **tropical cyclones**. As for all cyclones, tropical cyclones have low pressure in the cyclone center near sea level. Low-altitude winds also rotate cyclonically (counterclockwise in the N. Hemisphere) around these storms and spiral in towards their centers.

Tropical cyclones are called **hurricanes** over the Atlantic and eastern Pacific Oceans, the Caribbean Sea, and the Gulf of Mexico (Fig. 16.1). They are called **typhoons** over the western Pacific. Over the Indian Ocean and near Australia they are called **cyclones**. In this chapter we use "tropical cyclone" to refer to such storms anywhere in the world.

Comparing tropical and extratropical cyclones, tropical cyclones do not have fronts while mid-latitude cyclones do. Also, tropical cyclones have warm cores while mid-latitude cyclones have cold cores. Tropical cyclones can persist two to three times longer than typical mid-latitude cyclones. To help explain this behavior, we start by describing tropical cyclone structure.

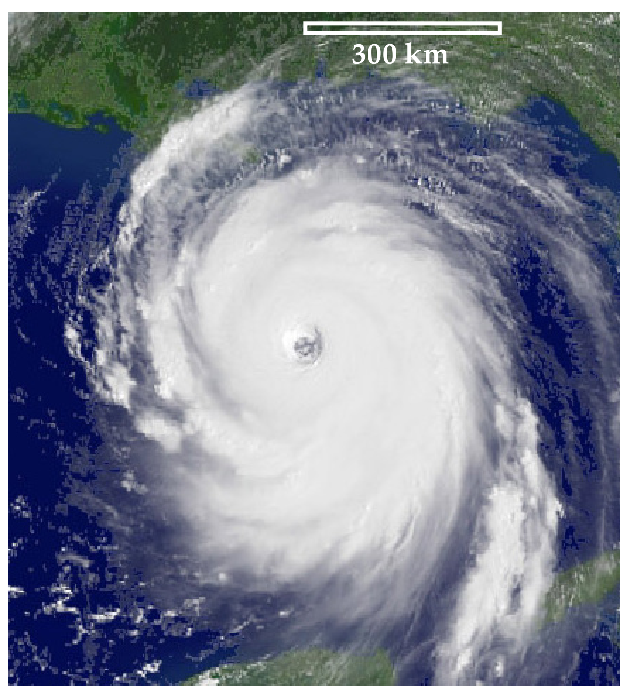


Figure 16.1
Visible-spectrum satellite picture of Hurricane Katrina over the Gulf of Mexico, taken 28 Aug 2005 at 1545 UTC. (GOES image courtesy of US DOC/NOAA.)

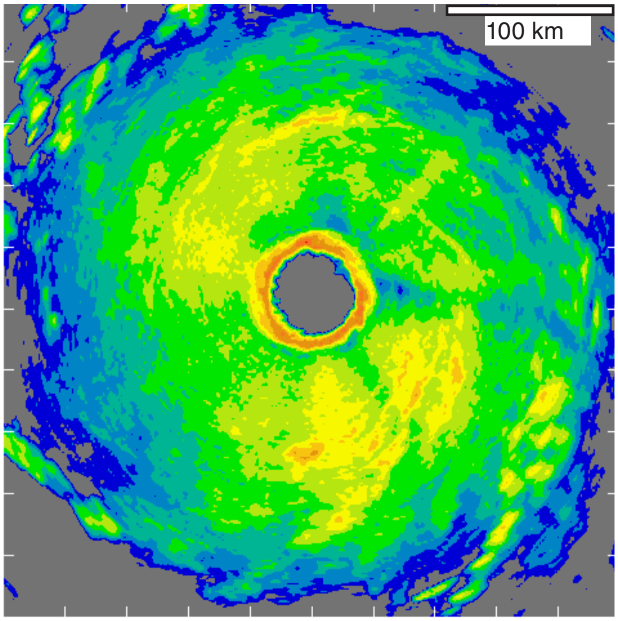


Figure 16.2
Radar reflectivity PPI image of Hurricane Katrina, taken from a research aircraft at altitude 2325 m on 28 Aug 2005 at 1747 UTC while it was over the Gulf of Mexico. Dark orange regions in the eyewall correspond to 35 to 40 dBZ radar reflectivity. (Image courtesy of US DOC/NOAA/AOML/ Hurricane Research Division.)

16.1. TROPICAL CYCLONE STRUCTURE

Tropical cyclones are made of thunderstorms. Near the center (**core**) of the tropical cyclone is a ring or circle of thunderstorms called the **eyewall**. This is the most violent part of the storm with the heaviest rain and the greatest radar reflectivity (Fig. 16.2). The tops of these thunderstorms can be in the lower stratosphere: 15 to 18 km high. Thunderstorm bases are very low: in the boundary layer. Thus, Tropical cyclones span the tropical-troposphere depth.

The anvils from each of the thunderstorms in the eyewall merge into one large roughly-circular **cloud shield** that is visible by satellite (Fig. 16.1). These anvils spread outward 75 to 150 km away from the eye wall. Hence, tropical cyclone diameters are roughly 10 to 20 times their depth (Fig. 16.3), although the high-altitude outflow from the top of some asymmetric tropical cyclones can reach 1000s of km (Fig. 16.4).

In the middle of the eyewall is a calmer region called the **eye** with warm temperatures, subsiding (sinking) air, and fewer or no clouds. Eye diameter at sea level is 20 to 50 km. The eye is conical, with the larger diameter at the storm top (Fig. 16.5).

Spiraling out from the eye wall can be zero or more bands of thunderstorms called **spiral bands**.

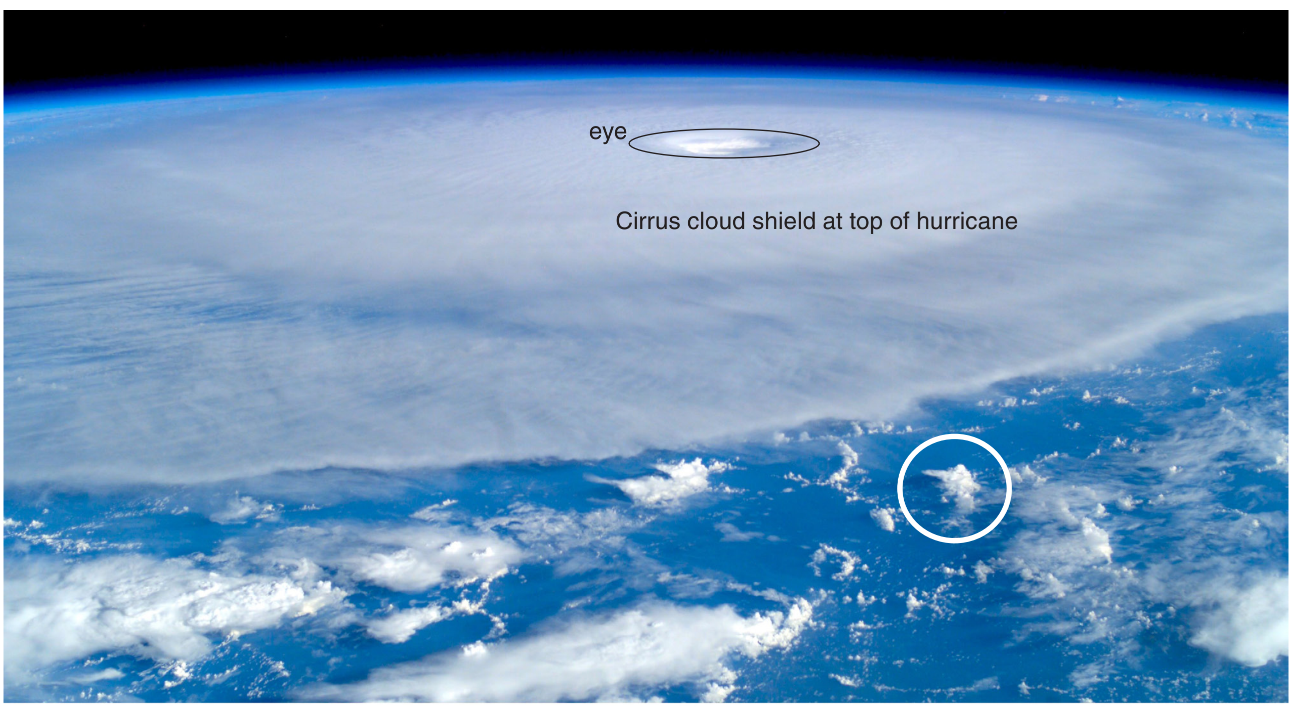


Figure 16.3
Hurricane Isabel as viewed from the International Space Station on 13 Sep 2003. The hurricane eye is outlined with the thin black oval, while the outflow from the eyewall thunderstorms fills most of this image. For scale comparison, the white circle outlines a nearby thunderstorm of depth and diameter of roughly 15 km. [Image courtesy of NASA - Johnson Spaceflight Center.]

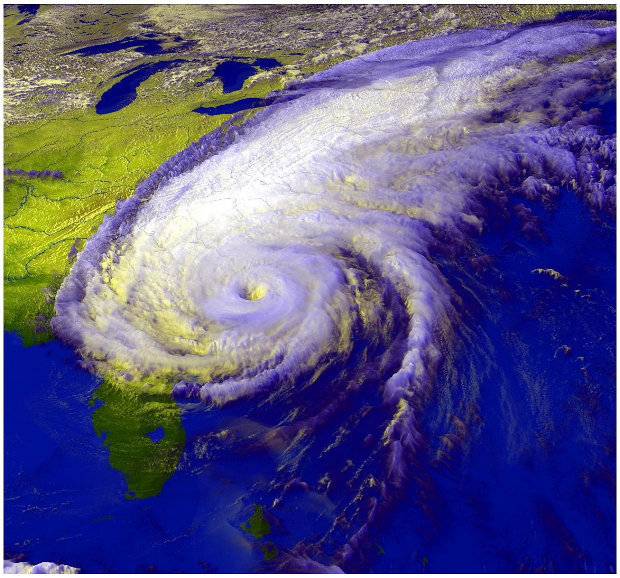


Figure 16.4
NOAA GOES satellite-derived 3-D rendering of hurricane Floyd at 2015 UTC on 15 Sep 1999, showing asymmetric outflow of this Category 4 storm extending 1000s of km north, past the Great Lakes into Canada. The eye is in the NW Atlantic just to the east of the Georgia-Florida border, USA. [Image by Hal Pierce is courtesy of the Laboratory for Atmospheres, NASA Goddard Space Flight Center.]

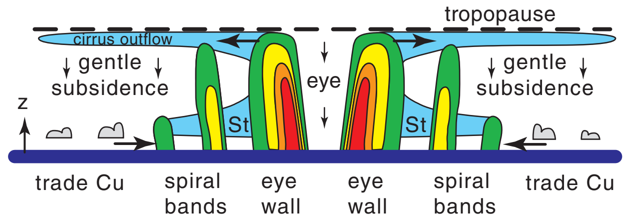


Figure 16.5
Vertical slice through a tropical cyclone. Green, yellow, orange, and red colors suggest the moderate to heavy rainfall as seen by radar. Light blue represents stratiform (St) clouds. The lighter rain that falls from the lower stratiform cloud deck of real tropical cyclones is not shown here. Tropopause altitude ≈ 15 km. Tropical-cyclone width ≈ 1500 km. Fair-weather trade-wind cumulus (Cu) clouds are sketched with grey shading. Arrows show radial and vertical wind directions.

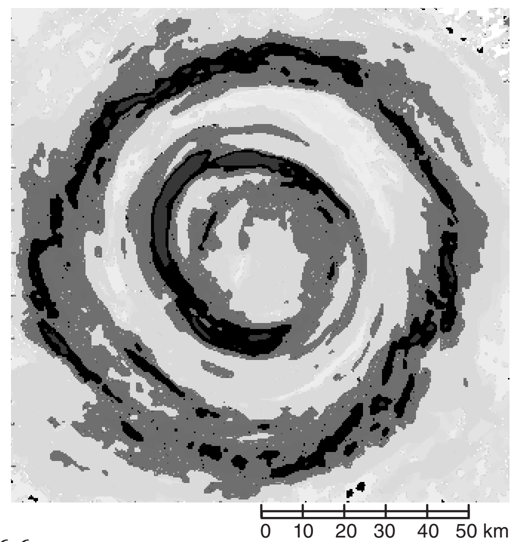


Figure 16.6
Zoomed view of double eyewall in Hurricane Rita, as viewed by airborne radar at 1 km altitude on 22 Sep 2005 at 1801 UTC. Darker greys show two concentric rings of heavier rain. Darkest grey corresponds to roughly 40 dBZ. [Modified from original image by Michael Bell and Wen-Chau Lee.]

Sometimes these spiral **rain bands** will merge to form a temporary second eyewall of thunderstorms around the original eye wall (Fig. 16.6). The lighter rain between the two eyewalls is called the **moat**.

During an **eyewall replacement cycle**, in very strong tropical cyclones, the inner eyewall dissipates and is replaced by the outer eyewall. When this happens, tropical cyclone intensity sometimes diminishes temporarily, and then strengthens again when the new outer eyewall diameter shrinks to that of the original eyewall.

16.2. INTENSITY & GEOGRAPHIC DISTRIBUTION

Lower sea-level pressures in the eye and faster winds in the lower troposphere indicate stronger tropical cyclones. Several different tropical cyclone scales have been devised to classify tropical cyclone strength, as summarized below.

16.2.1. Saffir-Simpson Hurricane Wind Scale

Herbert Saffir and Robert Simpson created a scale for Atlantic hurricane strength that came to be known as the **Saffir-Simpson Hurricane Scale**. This scale has been used since the 1970s to inform

Table 16-1. Saffir-Simpson Hurricane Wind Scale — Definition. Based on max 1-minute sustained wind speed measured at standard anemometer height of 10 m. Applies in the Atlantic Ocean, and in the eastern Pacific from the coast of the Americas to 180°W.

Category	Wind Speed			
	m s^{-1}	km h^{-1}	knots	miles h^{-1}
1	33 - 42	119 - 153	64 - 82	74 - 95
2	43 - 49	154 - 177	83 - 95	96 - 110
3	50 - 58	178 - 209	96 - 112	111 - 129
4	59 - 69	209 - 251	113 - 136	130 - 156
5	≥ 70	≥ 252	≥ 137	≥ 157

Table 16-2. Saffir-Simpson Hurricane Wind Scale — Description of Expected Damage — *Examples.*

Category	Concise Statement / Damage Expected / Examples
1	**Very dangerous winds will produce some damage.**
	People, livestock, and pets struck by flying or falling debris could be injured or killed. Older (mainly pre-1994 construction) mobile homes could be destroyed, especially if they are not anchored properly as they tend to shift or roll off their foundations. Newer mobile homes that are anchored properly can sustain damage involving the removal of shingle or metal roof coverings, and loss of vinyl siding, as well as damage to carports, sunrooms, or lanais. Some poorly constructed frame homes can experience major damage, involving loss of the roof covering and damage to gable ends as well as the removal of porch coverings and awnings. Unprotected windows may break if struck by flying debris. Masonry chimneys can be toppled. Well-constructed frame homes could have damage to roof shingles, vinyl siding, soffit panels, and gutters. Failure of aluminum, screened-in swimming pool enclosures can occur. Some apartment building and shopping center roof coverings could be partially removed. Industrial buildings can lose roofing and siding especially from windward corners, rakes, and eaves. Failures to overhead doors and unprotected windows will be common. Windows in high-rise buildings can be broken by flying debris. Falling and broken glass will pose a significant danger even after the storm. There will be occasional damage to commercial signage, fences, and canopies. Large branches of trees will snap and shallow rooted trees can be toppled. Extensive damage to power lines and poles will likely result in power outages that could last a few to several days.
	Hurricane Dolly (2008) brought Category 1 winds and impacts to South Padre Island, Texas.
2	**Extremely dangerous winds will cause extensive damage.**
	There is a substantial risk of injury or death to people, livestock, and pets due to flying and falling debris. Older (mainly pre-1994 construction) mobile homes have a very high chance of being destroyed and the flying debris generated can shred nearby mobile homes. Newer mobile homes can also be destroyed. Poorly constructed frame homes have a high chance of having their roof structures removed especially if they are not anchored properly. Unprotected windows will have a high probability of being broken by flying debris. Well-constructed frame homes could sustain major roof and siding damage. Failure of aluminum, screened-in swimming pool enclosures will be common. There will be a substantial percentage of roof and siding damage to apartment buildings and industrial buildings. Unreinforced masonry walls can collapse. Windows in high-rise buildings can be broken by flying debris. Falling and broken glass will pose a significant danger even after the storm. Commercial signage, fences, and canopies will be damaged and often destroyed. Many shallowly rooted trees will be snapped or uprooted and block numerous roads. Near-total power loss is expected with outages that could last from several days to weeks. Potable water could become scarce as filtration systems begin to fail.
	Hurricane Frances (2004) brought Category 2 winds and impacts to coastal portions of Port St. Lucie, Florida.
3	**Devastating damage will occur.**
	There is a high risk of injury or death to people, livestock, and pets due to flying and falling debris. Nearly all older (pre-1994) mobile homes will be destroyed. Most newer mobile homes will sustain severe damage with potential for complete roof failure and wall collapse. Poorly constructed frame homes can be destroyed by the removal of the roof and exterior walls. Unprotected windows will be broken by flying debris. Well-built frame homes can experience major damage involving the removal of roof decking and gable ends. There will be a high percentage of roof covering and siding damage to apartment buildings and industrial buildings. Isolated structural damage to wood or steel framing can occur. Complete failure of older metal buildings is possible, and older unreinforced masonry buildings can collapse. Numerous windows will be blown out of high-rise buildings resulting in falling glass, which will pose a threat for days to weeks after the storm. Most commercial signage, fences, and canopies will be destroyed. Many trees will be snapped or uprooted, blocking numerous roads. Electricity and potable water will be unavailable for several days to a few weeks after the storm passes.
	Hurricane Ivan (2004) brought Category 3 winds and impacts to coastal portions of Gulf Shores, Alabama.
4	**Catastrophic damage will occur.**
	There is a very high risk of injury or death to people, livestock, and pets due to flying and falling debris. Nearly all older (pre-1994) mobile homes will be destroyed. A high percentage of newer mobile homes also will be destroyed. Poorly constructed homes can sustain complete collapse of all walls as well as the loss of the roof structure. Well-built homes also can sustain severe damage with loss of most of the roof structure and/or some exterior walls. Extensive damage to roof coverings, windows, and doors will occur. Large amounts of windborne debris will be lofted into the air. Windborne debris damage will break most unprotected windows and penetrate some protected windows. There will be a high percentage of structural damage to the top floors of apartment buildings. Steel frames in older industrial buildings can collapse. There will be a high percentage of collapse to older unreinforced masonry buildings. Most windows will be blown out of high-rise buildings resulting in falling glass, which will pose a threat for days to weeks after the storm. Nearly all commercial signage, fences, and canopies will be destroyed. Most trees will be snapped or uprooted and power poles downed. Fallen trees and power poles will isolate residential areas. Power outages will last for weeks to possibly months. Long-term potable-water shortages will increase human suffering. Most of the area will be uninhabitable for weeks or months.
	Hurricane Charley (2004) brought Category 4 winds and impacts to coastal portions of Punta Gorda, Florida.
5	**Catastrophic damage will occur.**
	People, livestock, and pets are at very high risk of injury or death from flying or falling debris, even if indoors in mobile homes or framed homes. Almost complete destruction of all mobile homes will occur, regardless of age or construction. A high percentage of frame homes will be destroyed, with total roof failure and wall collapse. Extensive damage to roof covers, windows, and doors will occur. Large amounts of windborne debris will be lofted into the air. Windborne debris damage will occur to nearly all unprotected windows and many protected windows. Significant damage to wood roof commercial buildings will occur due to loss of roof sheathing. Complete collapse of many older metal buildings can occur. Most unreinforced masonry walls will fail, which can lead to collapse of buildings. A high percentage of industrial buildings and low-rise apartment buildings will be destroyed. Nearly all windows will be blown out of high-rise buildings resulting in falling glass, which will pose a threat for days to weeks after the storm. Nearly all commercial signage, fences, and canopies will be destroyed. Nearly all trees will be snapped or uprooted and power poles downed. Fallen trees and power poles will isolate residential areas. Power outages will last for weeks to possibly months. Long-term potable-water shortages will increase human suffering. Most of the area will be uninhabitable for weeks or months.
	Hurricane Andrew (1992) brought Category 5 winds and impacts to coastal portions of Cutler Ridge, Florida.

From an online report "The Saffir-Simpson Hurricane Wind Scale", National Hurricane Center, National Weather Service, NOAA. 2010.

disaster-response officials about hurricane strength.

In 2010, the US National Hurricane Center updated this scale (now called the **Saffir-Simpson Hurricane Wind Scale**), and defined hurricane intensity on wind speed only (Table 16-1). The scale ranges from category 1 for a weak hurricane to category 5 for a strong hurricane. Table 16-2 has a description of the expected damage for each category.

16.2.2. Typhoon Intensity Scales

For typhoons in the western North Pacific Ocean, storm intensity has been classified three different ways by three different organizations: the **Japan Meteorological Agency (JMA)**, the **Hong Kong Observatory (HKO)**, and the **US Joint Typhoon Warning Center (JTWC)**. See Table 16-3.

16.2.3. Other Tropical-Cyclone Scales

Additional tropical-cyclone intensity scales (with different max wind-speed M_{max} definitions and category names) have been defined by agencies in:

- Australia (**Australian Bureau of Meteorology**, for S. Hemisphere east of 90°E). Cat. 1 for $M_{max} < 125$; Cat. 2 for $125 \le M_{max} \le 164$; Cat. 3 for $165 \le M_{max} \le 224$; Cat. 4 for $225 \le M_{max} \le 279$; and Cat. 5 for $280 < M_{max}$, where all winds are in km h^{-1}.
- India (**Regional Specialized Meteorological Center in New Delhi**, for N. Hemisphere Indian Ocean between 45°E and 100°E).
- France (**Météo-France**, for S. Hemisphere west of 90°E).

Even more confusing, some agencies use 1-minute average (sustained) winds, some use 10-minute average (sustained) winds, and some use gusts.

16.2.4. Geographic Distribution and Movement

Fig. 16.7 outlines the regions of greatest frequency of tropical cyclones and shows typical storm tracks.

Tropical cyclones are steered mostly by the large-scale global circulation. Most tropical cyclones form between 10° and 30° latitude, which is the trade-wind region. Hence, most tropical cyclones are steered from east to west initially.

Later, these storms tend to turn poleward under the influence of monsoon circulations. For example, the Bermuda High (also known as the Azores High) over the North Atlantic Ocean has a clockwise circulation around it (see Fig. 11.31 in the General Circulation chapter), which turns tropical cyclones northward in the Western North Atlantic Ocean. The tracks of tropical cyclones and storms that have weakened from tropical cyclones to tropical storms

INFO • Cyclone Damage Potential (CDP)

The total amount of destruction depends on the <u>area</u> of hurricane force winds, max wind <u>speed</u>, and the storm <u>duration</u> over any location. For this reason, the insurance industry utilizes a **<u>cyclone damage potential index</u>** (CDP) defined as follows:

$$CDP = \left(\frac{M_{to}}{M_t}\right) \cdot \left[\left(\frac{M_{\max}}{M_{\max o}}\right)^3 + 5\left(\frac{R_h}{R_{ho}}\right) \right]$$

where M_t = translation speed (km h^{-1}) of the eye, M_{max} is max surface wind speed (km h^{-1}), R_h = average radius (km) of hurricane force winds, and the corresponding parameters are M_{to} = 7.4 km h^{-1}, M_{maxo} = 120 km h^{-1}, R_{ho} = 96 km. If $M_t < 9.25$ km h^{-1}, then set $M_t = 9.25$ km h^{-1}. CDP varies between 0 and 10 (dimensionless), with 10 indicating the worst destruction.

For example, when hurricane Katrina was a category 5 storm before landfall, it had M_t = 20 km h^{-1}, M_{max} =280 km h^{-1}, & R_h = 100 km, yielding **<u>CDP ≈ 6.7</u>** .

Table 16-3. Typhoon (tropical cyclone) intensity scale, based on max sustained winds during 10-minute periods. Defined by JMA, unless otherwise noted. Applies between 100°E to 180°E.

Category	Wind Speed m s^{-1}	Wind Speed km h^{-1}	Wind Speed knots
Tropical Depression	< 17	< 61	< 33
Tropical Storm	17 - 24	62 - 88	34 - 47
Severe Tropical Storm	25 - 32	89 - 117	48 - 63
Typhoon	33 - 41	118- 56	64 - 74
Severe (or Very Strong) Typhoon	42 - 51 (HKO) 44 - 54	150 - 184 (HKO) 157 - 193	80 - 99 (HKO) 85 - 104
Super Typhoon	> 51 (HKO) > 59 (JTWC)	>185 (HKO) > 211 (JTWC) > 194	>100 (HKO) > 114 (JTWC) > 105

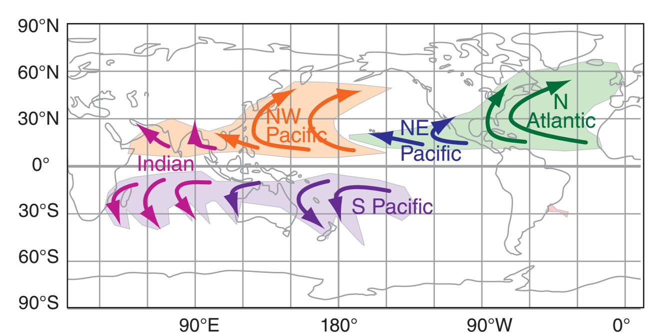
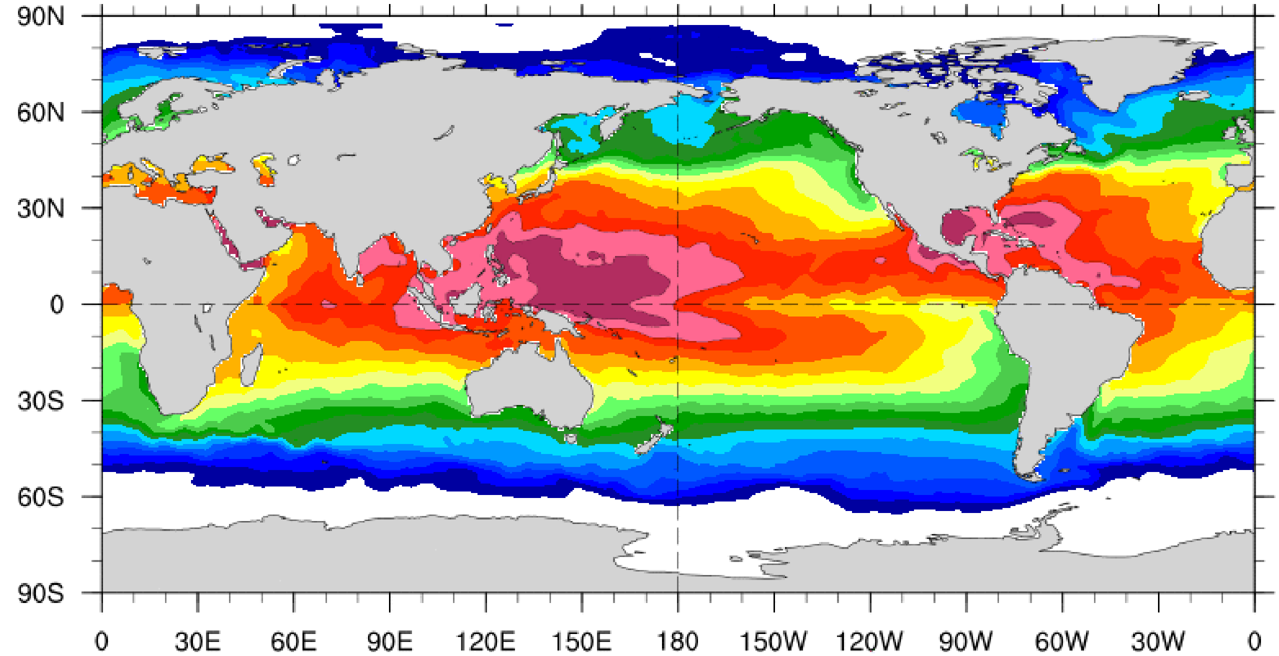


Figure 16.7
Map of tropical-cyclone locations (shading), with a general sense of the cyclone tracks (arrows).



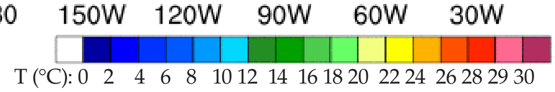


Figure 16.8
Sea surface temperatures (°C) as one-week average during 4 - 10 Sep 2016. [Courtesy of the NOAA/Earth System Research Lab/Physical Sciences Division.]

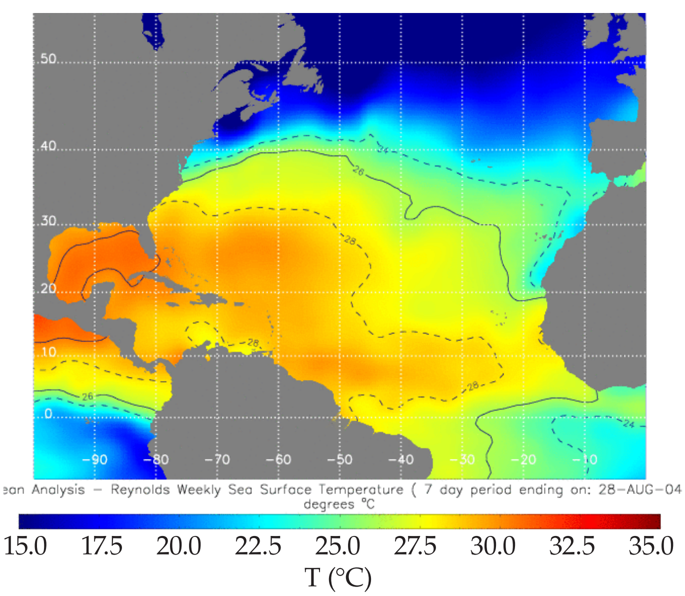


Figure 16.9
Analysis of Atlantic-ocean sea-surface temperatures (°C) averaged over 7 days ending on 28 Aug 2004. Hurricane Francis occurred during this time period . [Adapted from a USA National Hurricane Center image using NCEP/NOAA data.]

continue to turn toward the northeast as they proceed further into the mid-latitudes and encounter west winds in the global circulation.

A striking observation in Fig. 16.7 is that no tropical cyclones form at the equator. Also, none cross the equator. The next section explains why.

16.3. EVOLUTION

16.3.1. Requirements for Cyclogenesis

Seven conditions are necessary for tropical cyclones to form: a warm sea surface, non-zero Coriolis force, nonlocal conditional instability, high humidity in the mid troposphere, weak ambient wind shear, enhanced synoptic-scale vorticity, and a trigger.

16.3.1.1. Warm Sea Surface

The **sea surface temperature** (**SST**) must be approximately 26.5°C or warmer (Figs. 16.8 and 16.9 on this page, and 16.51 at the end of this chapter), and the warm surface waters must be at least 50 m deep. This warm temperature is needed to enable strong evaporation and heat transfer from the sea surface into the boundary layer. The warmer, more-humid boundary-layer air serves as the fuel for thunderstorms in the tropical cyclone.

The fast winds in tropical cyclones create large waves that stir the top part of the ocean. If the warm waters are too shallow, then this turbulent mixing will stir colder deeper water up to the surface (see INFO Box two pages later). When this happens, the sea-surface temperature decreases below the 26.5°C threshold, and the tropical cyclone kills itself.

16.3.1.2. Coriolis Force

Tropical cyclones cannot exist within about 500 km of the equator (i.e., ≤ 5° latitude), because Coriolis force is near zero there (and is exactly zero right at the equator). Rarely, very small-diameter tropical cyclones have been observed closer to the equator, but none are observed right at the equator. Tropical cyclones cannot form at the equator, and existing cyclones cannot cross the equator (Fig. 16.7).

Without Coriolis force, boundary-layer air would be sucked directly into the eye by the low pressure there. Thus, air molecules would accumulate in the eye, causing pressure to increase towards ambient values. The result is that the low pressure would disappear, winds would die, and the tropical cyclone would cease to exist in less than a day. This is what happens to tropical cyclones that approach the equator (Fig. 16.10).

But with Coriolis force, winds in the bottom half of the troposphere are forced around the eye at gradient- or cyclostrophic-wind speeds. Namely, most of the air is flowing tangentially around the eye rather than flowing radially into it. Only close to the ground does drag change the winds into boundary-layer gradient winds, resulting in a small amount of convergence toward the eye. Thus, the tropical cyclone can persist for many days.

16.3.1.3. Nonlocal Conditional Instability

Because tropical cyclones are made of thunderstorms, the tropical environment must have sufficient nonlocal conditional instability to support deep thunderstorm convection. Namely, there must be a stable layer (i.e., a cap) above a warm humid boundary layer, and the mid-troposphere must be relatively cool compared to the boundary layer.

The warm humid boundary layer is achieved via strong heat and moisture fluxes from the warm sea surface into the air. The cap is the trade-wind inversion that was discussed in the General Circulation chapter. These conditions, combined with a cool mid troposphere, lead to large values of convective available potential energy (CAPE), as was thoroughly explained in the Thunderstorm chapter. Hence, large values of CAPE imply sufficient nonlocal conditional instability for tropical cyclones.

16.3.1.4. High Humidity in Mid-troposphere

In a deep layer of air centered at roughly 5 km above sea level, humidity must be high. Otherwise,

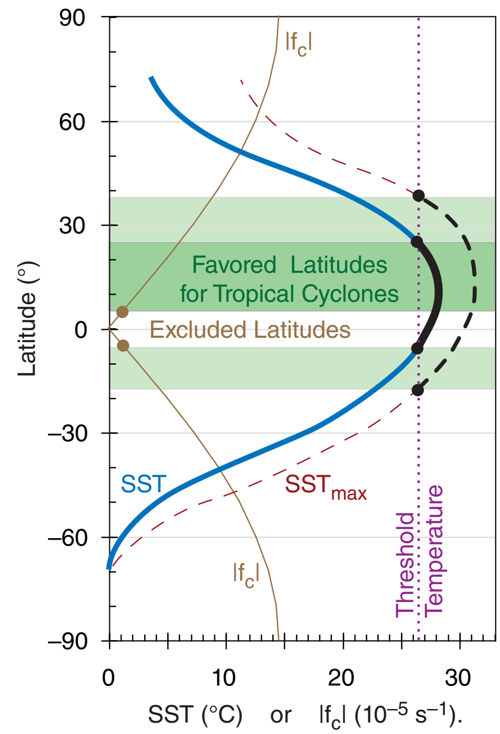


Figure 16.10
Approximate zonal average sea-surface temperature SST (°C, thick solid blue line) averaged around the globe on 3 Aug 2009. Also shown is the max SST (°C, thin dashed red line) on that day at each latitude. The magnitude of the Coriolis parameter f_c (10^{-5} s^{-1}) is indicated by the thin solid brown line. Dotted purple line indicates the SST threshold of roughly 26.5°C for tropical cyclones. Those latitudes that exceed the temperature threshold and have Coriolis parameter that is not near zero are favored locations for tropical cyclones (shaded medium green). Tropical cyclones can also occur in limited regions (shaded light green) at other latitudes where the max SST exceeds the threshold. "Excluded latitudes" are those for which Coriolis force is too small. As the seasons progress, the location of peak SST shifts, causing the favored latitudes to differ from this particular August example.

the incipient thunderstorms cannot continue to grow and organize into tropical cyclones.

Note that this differs from mid-latitude thunderstorms, where a drier mid-troposphere allows more violent thunderstorms. When dry environmental air is entrained into the sides of mid-latitude thunderstorms, some of the storm's hydrometeors evaporate, causing the strong downdrafts that define supercell storms and which create downbursts, gust fronts, and can help trigger tornadoes.

In the tropics, such downbursts from any one thunderstorm can disrupt neighboring thunderstorms, reducing the chance that neighboring thunderstorms can work together in an incipient eyewall. Also, the cold downburst air accumulates at the bottom of the troposphere, thereby increasing static stability and reducing deep convection.

INFO • Trop. Cyclone-induced Currents

Near-surface winds cause net ocean transport 90° to the right of the wind vector in the N. Hemisphere, as explained later in the Storm Surge section. Given tangential winds that circle a tropical cyclone, we expect the net Ekman transport (average ocean currents induced by the tropical cyclone) to be outward (Fig. 16.a)

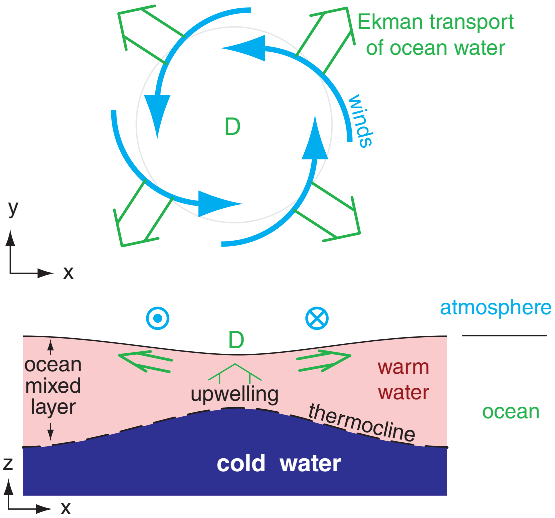


Fig. 16.a. *Ocean currents (green double-shaft arrows) induced by tropical-cyclone winds (cyan arrows).*

Such horizontal divergence (D) of the ocean-surface water does three things: (1) it lowers the sea surface by removing sea water; (2) it causes upwelling of water toward the surface; and (3) it causes the colder deep water to be brought closer to the surface (Fig. 16.a-bottom). The **thermocline** is the interface between cold deep water and warmer mixed-layer above.

The tropical cyclone brings cold water closer to the surface, where ocean mixed-layer turbulence caused by breaking waves can further mix the warm and cold waters. Thus, tropical cyclones will kill themselves by cooling the sea-surface temperature unless the pre-storm warm ocean mixed-layer is ≥ 50 m deep.

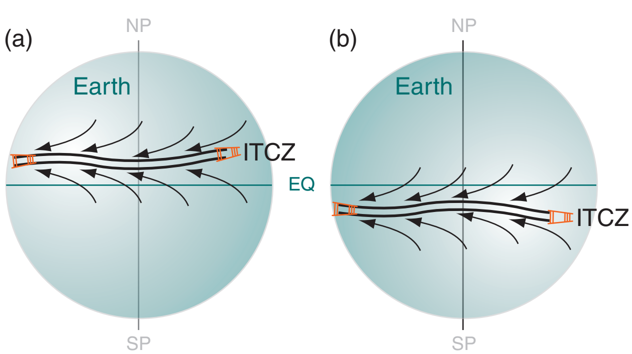


Figure 16.11. *Sketch of the Earth showing the Intertropical Convergence Zone (ITCZ, double solid line) for the early Autumn tropical-cyclone seasons during (a) August and (b) February. NP = North Pole. SP = South Pole. EQ = equator.*

16.3.1.5. Weak Ambient Wind Shear

Wind shear within four degrees of latitude of the incipient storm must be weak (ΔM < 10 m s^{-1} between pressure levels 80 and 25 kPa) to enable thunderstorm clusters to form. These clusters are the precursors to tropical cyclones.

If the shear is too strong, the updrafts in the thunderstorms become tilted, and latent heating due to water-vapor condensation is spread over a broader area. This results in less-concentrated warming, and a reduced ability to create a low-pressure center at sea level around which the thunderstorms can become organized into a tropical cyclone.

This requirement differs from that for mid-latitude thunderstorms. At mid-latitudes, strong shear in the ambient environment is good for storms because it encourages the creation of mesocyclones and supercell thunderstorms. Apparently, in the tropics such rotation of individual thunderstorms is bad because it interferes with the collaboration of many thunderstorms to create an eyewall.

16.3.1.6. Synoptic-scale Vorticity

A relative maximum of relative vorticity in the bottom half of the troposphere can help organize the thunderstorms into an incipient tropical cyclone. Otherwise, any thunderstorms that form would act somewhat independently of each other.

16.3.2. Tropical Cyclone Triggers

Even if all six previous conditions are satisfied, a method to trigger the tropical cyclone is also needed. A trigger is anything that creates synoptic-scale horizontal convergence in the atmospheric boundary layer. Such horizontal convergence forces upward motion out of the boundary-layer top as required by mass conservation (see the Atmospheric Forces & Winds chapter). Synoptic-scale upward motion can initiate and support an organized cluster of thunderstorms — incipient tropical cyclones.

Some triggers are: the ITCZ, easterly waves, Monsoon troughs, mid-latitude fronts that reach the tropics, and Tropical Upper Tropospheric Troughs.

16.3.2.1. ITCZ

The **Intertropical Convergence Zone** (**ITCZ**, see General Circulation chapter) is the region where the northeasterly trade winds from the Northern Hemisphere meet the southeasterly trade winds from the Southern Hemisphere. This convergence zone shifts between about 10°N during August and September (end of N. Hemisphere summer, Fig. 16.11), and 10°S during February and March (end of S. Hemisphere summer). During these months of maximum asymmetry in global heating, the ITCZ is far enough from the equator for sufficient Coriolis force to enable tropical cyclones.

16.3.2.2. African Easterly Wave

Although equatorial Africa is hot, the Sahara Desert (roughly spanning 15° to 30°N) and Arabian Peninsula are even hotter. This temperature excess is within an atmospheric boundary layer that can reach 5 km depth by late afternoon, although average depth over 24 h is about 3 km. Thus, along the southern edge of the Sahara and the Arabian Peninsula is a strong temperature gradient (Fig. 16.12).

As explained in the thermal-wind section of the General Circulation chapter, a north-south temperature gradient creates a north-south pressure gradient that increases with altitude (up to the top of the boundary layer for this scenario). For Africa, higher pressure aloft is over the Sahara Desert and lower pressure aloft is further south over the Sudan and Congo.

This pressure gradient drives a geostrophic wind from the east over the African Sahel (at about 15°N). These winds reach maximum speeds of 10 to 25 m s^{-1} at an altitude of about 3 km in late summer. This is called the **African Easterly Jet** (**AEJ**).

Like the subtropical and mid-latitude jets, the AEJ is barotropically and baroclinically unstable. This means that the jet tends to meander north and south while blowing from east to west. The mountains in eastern Africa (Ethiopia Highlands and Darfur Mountains) can trigger such oscillations.

The oscillations in this low-altitude jet are called waves, and are known as **Easterly Waves** because the wave troughs and crests move from the east at speeds of 7 to 8 m s^{-1}. These waves have wavelength of about 2,000 to 2,500 km, and have a wave period of 3 to 4 days. Although these waves are created over Africa, they continue to propagate west across the tropical North Atlantic, the Caribbean, central America, and finally into the tropical Northeast Pacific (Fig. 16.13).

Easterly waves can trigger tropical cyclones because of the low-altitude convergence that occurs east of the wave troughs. These waves are found predominantly in a latitude band between 10°N and 30°N, where there is lower pressure to the south and the subtropical high-pressure belt to the north. Thus, a trough of low pressure in an easterly wave corresponds to a region where the isobars meander to the north (Fig. 16.14) in the N. Hemisphere.

Because the pressure gradient is weak and sometimes difficult to analyze from sparsely spaced weather stations in this region, tropical meteorologists will often look at the streamlines instead of isobars. Streamlines are weather-map lines that are everywhere parallel to the flow at any instant in time

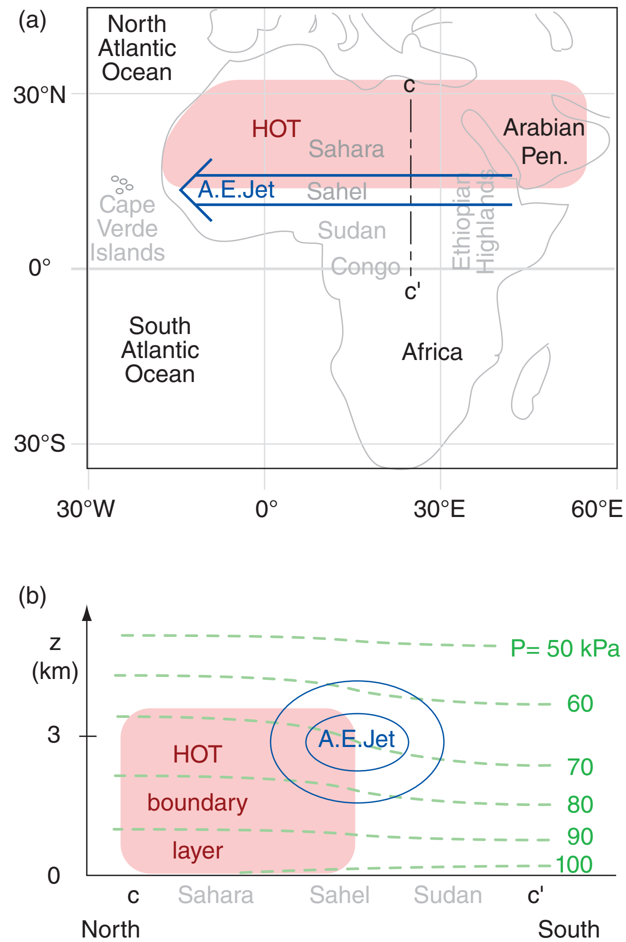


Figure 16.12
African Easterly (A.E.) Jet. (a) Map of Africa. (b) Vertical cross section along c - c' in (a), where thin blue ovals represent isotachs around the core of the AEJ. Pink area indicates hot air. Note the low altitude of this A.E. jet.

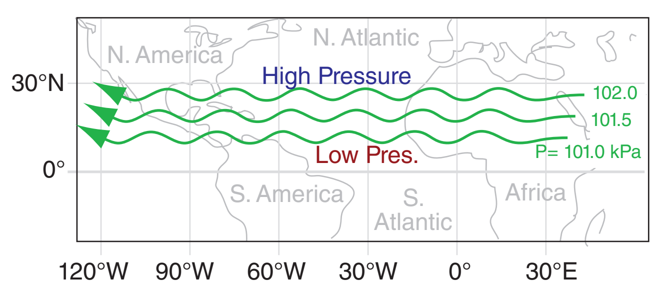


Figure 16.13
Easterly waves near the bottom of the troposphere. Wavelength is roughly 20° to 25° of longitude (≈ 2000 to 2500 km). Isobars of sea-level pressure are solid green lines with arrows. Wave troughs propagate toward the west. Streamlines are roughly parallel to the isobars.

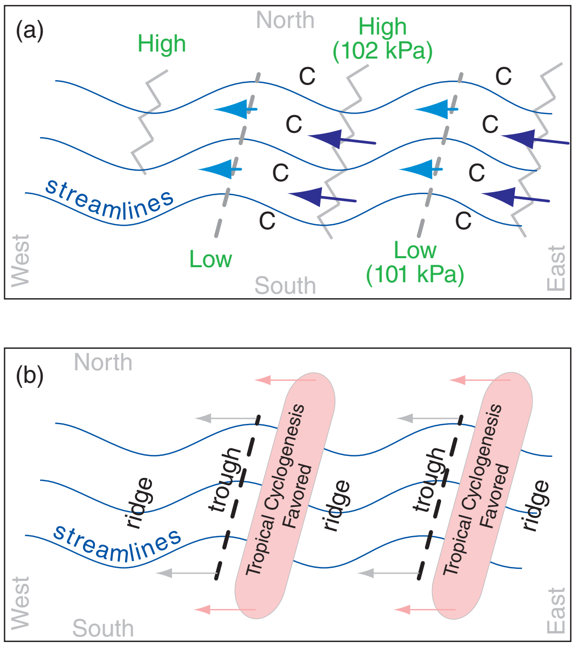


Figure 16.14
Characteristics of easterly waves in lower tropospheric flow over the tropical North Atlantic. (a) Blue arrows indicate winds; longer arrows denote faster winds; C = convergence. (b) Grey and pink arrows show how the easterly waves and the tropical cyclogenesis regions (pink shading) propagate toward the west.

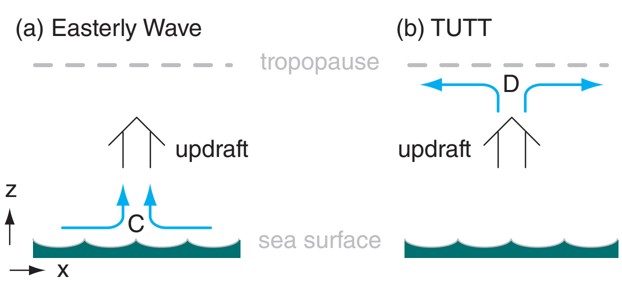


Figure 16.15
Mechanisms that create updrafts to trigger thunderstorm clusters: (a) C = lower-tropospheric horizontal convergence in an easterly wave; (b) D = horizontal divergence aloft in a Tropical Upper Tropospheric Trough (TUTT).

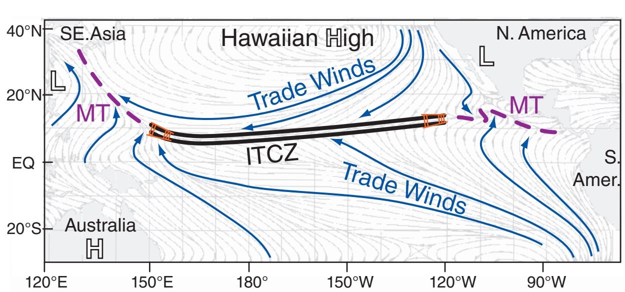


Figure 16.16
Streamline analysis (blue lines with arrows) of near-surface winds, averaged during August. MT denotes Monsoon Trough, shown by the thick dashed lines. 𝕃 and ℍ are monsoon lows and highs. [modified from Naval Research Lab. image]

(see the Regional Winds chapter). They more-or-less follow the isobars. Thus, regions where streamlines for low-altitude winds meander to the north can also be used to locate troughs in the easterly waves.

Recall that mid-latitude cyclones are favored in the updrafts under divergence regions in the polar jet (centered at the top of the troposphere). However, for tropical cyclogenesis we also look for updrafts above the convergence regions of the jet near the bottom of the troposphere (Fig. 16.15a).

Where are these convergence regions in easterly waves? According to the Atmospheric Forces & Winds chapter, winds are slower-than-geostrophic around troughs, and faster around ridges. This rule continues to hold in the tropics and subtropics. Thus, the convergence zone with fast inflow and slow outflow is east of the trough axis (Fig. 16.14a).

Also, since the trough axis often tilts eastward with increasing altitude, it means that a mid-tropospheric trough is often over a lower-tropospheric convergence region. This trough has the cyclonic vorticity that encourages organization of thunderstorms into incipient tropical cyclones.

It is these regions east of troughs in easterly waves where the thunderstorms of incipient tropical cyclones can be triggered (Fig. 16.14b). As the easterly waves propagate from east to west, so move the convergence regions favoring tropical cyclone formation. This convergence mechanism is important for tropical cyclone triggering in both the North Atlantic and Northeast Pacific. Roughly 85% of intense Atlantic hurricanes are triggered by easterly waves.

16.3.2.3. Monsoon Trough

A **monsoon trough** (MT) is where the ITCZ merges into a monsoon circulation. For example, Fig. 16.16 shows a monsoon low pressure (𝕃) region over southeast Asia (left side of Fig.), with winds rotating counterclockwise around it. Near the dashed line labeled MT in the western Pacific, monsoon winds from the southwest converge with easterly trade winds from the Hawaiian (Pacific) High.

Thus, the monsoon trough in the western Pacific is a convergence region that can trigger thunderstorms for incipient typhoons. Similar monsoon troughs have been observed in the Indian Ocean and the Eastern Pacific. In Fig. 16.16, the region where the monsoon trough in the western Pacific meets the ITCZ has enhanced convergence, and is a more effective trigger for tropical cyclones.

16.3.2.4. Mid-latitude Frontal Boundary

Sometimes cold fronts from midlatitudes can move sufficiently far equatorward to reach the subtropics. Examples include cold fronts reaching southern Florida (Fig. 16.17). Although these fronts are often weak and slow moving by the time the reach the subtropics, they nonetheless have convergence across them that can trigger tropical cyclone thunderstorms in the Caribbean and the subtropical western North Atlantic.

16.3.2.5. TUTT

TUTT stands for **Tropical Upper Tropospheric Trough**. This is a high-altitude cold-core low-pressure system that can form in the subtropics. East of the TUTT axis is a region of horizontal divergence aloft, which creates upward motion below it in order to satisfy air mass continuity (Fig. 16.15b). This upward motion can help trigger thunderstorms as precursors to tropical cyclones.

16.3.3. Life Cycle

At locations where all of the necessary conditions are met (including any one trigger), the incipient tropical cyclones usually progress through the following intensification stages: tropical disturbance, tropical depression, tropical storm, tropical cyclone.

16.3.3.1. Tropical Disturbance

The US National Hurricane Center defines a **tropical disturbance** as "a discrete tropical weather system of apparently organized convection (generally 200 to 600 km in diameter) originating in the tropics or subtropics, having a nonfrontal migratory character, and maintaining its identity for 24 hours or more." Namely, it is a cluster of thunderstorms that stay together as they move across the ocean.

This thunderstorm cluster is visible by satellite as consisting of distinct thunderstorms with their own anvils and separate precipitation regions. There is no eye, and little or no rotation visible. Some tropical disturbances form out of **Mesoscale Convective Systems** (**MCS**s), particularly those that develop a **Mesoscale Convective Vortex** (**MCV**, see the Thunderstorm chapter). Some are identified as the thunderstorm clusters that move with easterly waves.

Most tropical disturbances do not evolve into tropical cyclones. Hence, tropical disturbances are not usually named or numbered. However, tropical meteorologists watch them carefully as they evolve, as potential future tropical cyclones.

As condensation and precipitation continue in this thunderstorm cluster, more and more latent heat is released in those storms. Namely, the mid-tropospheric air near this cluster becomes warmer than the surrounding ambient air. According to the hypsometric relationship, pressure decreases more slowly with increasing altitude in this warm region. Thus, synoptic-scale high pressure starts to form near the top of the troposphere in the region of this cluster.

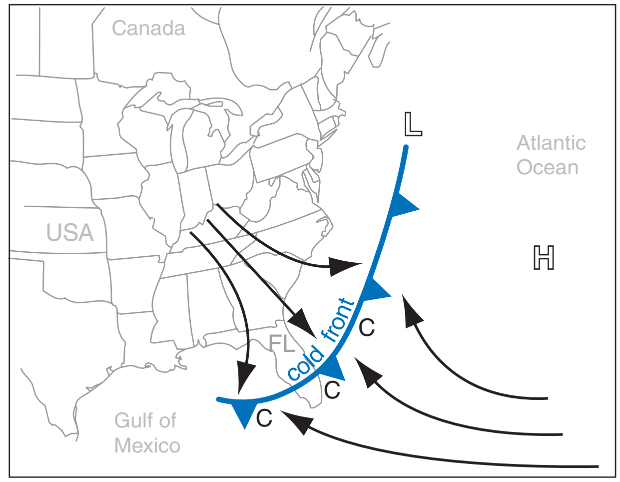


Figure 16.17
Lower tropospheric convergence (C) along a cold front over Florida (FL), USA.

16.3.3.2. Tropical Depression (TD)

The high pressure aloft begins to create a thermal circulation (see General Circulation chapter), where air aloft is driven horizontally outward — down the pressure gradient toward the lower pressure outside of the cluster. This diverging air aloft also starts to rotate anticyclonically.

The outward moving air aloft removes air molecules from the thunderstorm cluster region of the troposphere, thereby lowering the surface pressure under the cluster. The "depressed" pressure at sea level is why this stage gets its name: **tropical depression**. Tropical depressions are given an identification number, starting with number 1 each year in each region.

As this weak surface low forms, it creates a pressure gradient that starts to suck winds horizontally toward the center of the low. This inflowing air begins rotating cyclonically due to Coriolis force. The surface low pressure is usually too weak to measure directly. However, this tropical depression stage is defined by measurable near-surface winds of 17 m s^{-1} or less (Table 16-4) turning cyclonically as a closed circulation around the cluster. There is still no eye at this stage, however the thunderstorms are becoming aligned in spiral rain bands (Fig. 16.18). Most tropical depressions do not strengthen further.

However, for the few storms that do continue to strengthen, the near-surface winds are boundary-layer gradient winds that spiral in towards the center of the cluster. This radial inflow draws in more

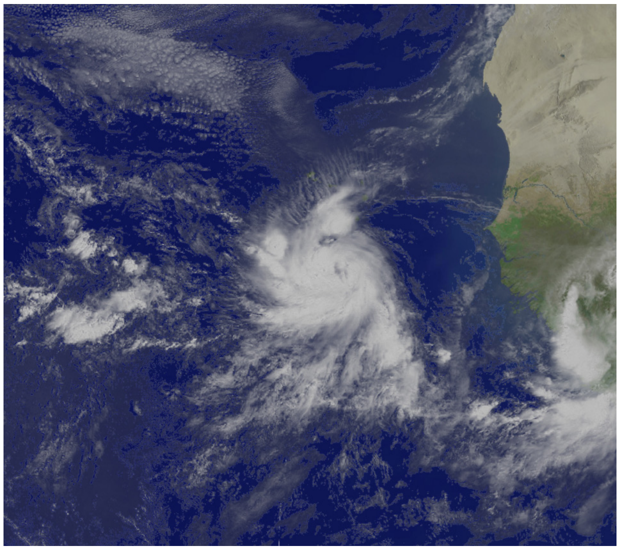


Figure 16.18
Tropical Depression 10 (in center of photo), over the tropical Atlantic Ocean south-southeast of the Cape Verde Islands. Africa is visible in the right quarter of the photo. (Visible satellite image taken 11:45 UTC on 2 Sep 2008, courtesy of NOAA.)

Table 16-4. **Stages leading to tropical cyclones.** M_{max} = near-surface wind-speed maximum. n/a = not applicable.			
Stage	**M$_{max}$** (m s^{-1})	(km h^{-1})	(knots)
Tropical Disturbance	n/a	n/a	n/a
Tropical Depression	< 17	< 61	≤ 33
Tropical Storm	17 - 32	61 - 118	33 - 63
Tropical Cycl.(Hurricane)	≥ 33	≥ 119	≥ 64

warm humid boundary-layer air, thereby refueling the thunderstorms and allowing them to persist and strengthen. Condensation and precipitation increase, causing a corresponding increase in latent-heat release.

This warm core of the storm further strengthens the high pressure aloft via the hypsometric relationship, which drives more outflow aloft and removes more air molecules from the troposphere near the cluster. Thus, the surface low pressure can continue to deepen in spite of the boundary-layer inflow. Namely, the high pressure aloft creates an exhaust system, while the boundary layer is the intake system to the incipient cyclone.

16.3.3.3. Tropical Storm (TS)

When the surface low is deep enough to drive winds faster than 17 m s^{-1} (but less than 33 m s^{-1}) in a closed cyclone circulation, then the system is classified as a **tropical storm**. On weather maps, it is indicated with symbol .

At this point, the thunderstorm rain bands have nearly completely wrapped into a circle — the future eyewall (Fig. 16.19d). The anvils from the thunderstorms have usually merged in to a somewhat circular **central dense overcast** (**CDO**), which is clearly visible as a large-diameter cold high cloud. There is no eye at this stage. Tropical storms are organized sufficiently to be able to modify their local environment to allow them to persist, without relying so much on a favorable ambient environment.

At this stage the storm is given a name for identification (and its former tropical-depression number is dropped). The same name is used if the storm further strengthens into a tropical cyclone, and for the rest of its life cycle.

Each ocean region has its own naming convention for these tropical storms. Hurricanes (Atlantic

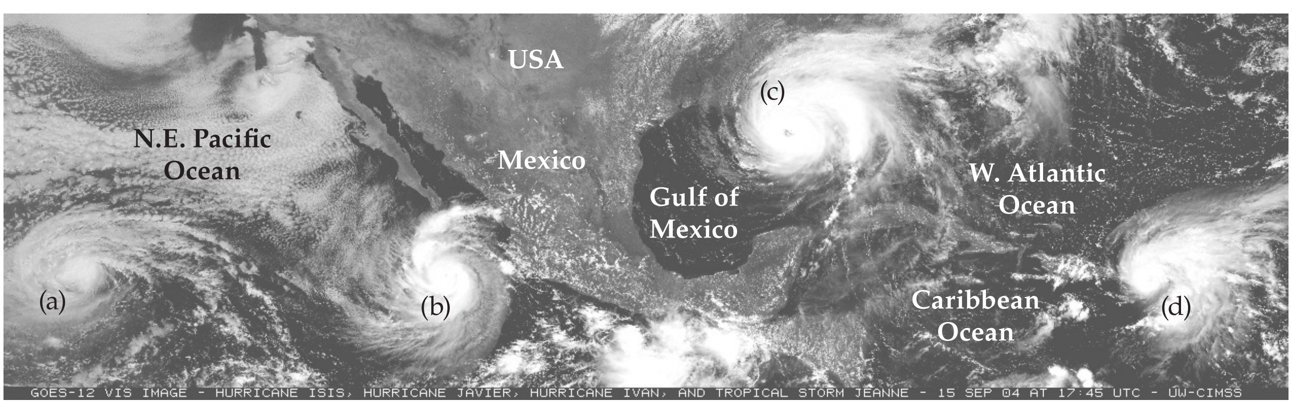


Figure 16.19
Visible satellite image at 17:45 UTC on 15 Sep 2004, showing four tropical systems. From left to right: (a) Hurricane Isis, (b) Hurricane Javier, (c) Hurricane Ivan, (d) Tropical Storm Jeanne. This image covers the eastern N. Pacific Ocean, Central America, Gulf of Mexico, Caribbean Ocean, and western N. Atlantic. (Image courtesy of University of Wisconsin–Madison CIMSS/ SSEC.)

and Eastern Pacific) and Cyclones (Indian Ocean and Southwestern Pacific near Australia) are usually given names of men or women. These are assigned in alphabetical order according to lists that have been set in advance by the **World Meteorological Organization** (**WMO**). Typhoons are given names of flowers, animals, birds, trees, foods (not in alphabetical order but in an order set by the WMO's Typhoon Committee).

Tropical-storm and tropical-cyclone names are re-used in a six-year cycle. However the names of the strongest, most destructive tropical cyclones are "retired" and never used again (they are replaced with other names starting with the same letter).

16.3.3.4. Tropical Cyclone (Hurricane, Typhoon)

Roughly half of the tropical storms in the Atlantic continue to strengthen into tropical cyclones. By this stage there is a well defined eye surrounded by an eyewall of thunderstorms. Max wind speeds are 33 m s^{-1} or greater in a closed circulation around the eye. The central dense overcast usually has a cloud-free eye. The weather map symbol for a tropical cyclone is .

There is heavy precipitation from the eyewall thunderstorms, and the core of the storm is significantly warmer than the environment outside the storm. Sea-level pressure decrease in the eye is measurable and significant, and a rise of sea level into a storm surge is possible. The storm organization allows it to persist for days to weeks as it moves across ocean basins (Fig. 16.19a-c).

16.3.4. Movement/Track

16.3.4.1. Steering by the General Circulation

Tropical cyclones are moved by the larger-scale winds near them. These winds are variable, causing a wide variety of hurricane tracks and translation speeds. However, by focusing on the climatological average trade winds and the monsoon circulations (Figs. 16.7 & 16.20), we can then anticipate average movement of tropical cyclones.

During late Summer and early Fall, monsoon high-pressure regions form over the oceans, as was discussed in the General Circulation chapter. Winds rotate clockwise (counterclockwise) around these highs in the Northern (Southern) Hemisphere. Equatorward of the high center the monsoon winds are in the same direction as the trade winds — from east to west. Thus, average tropical cyclone tracks in the eastern and central longitudes of an ocean basin are usually zonal — from east to west.

However, near the west sides of the ocean basins, the monsoon winds turn poleward (Fig. 16.20), and thus steer the tropical cyclones away from the equator. Many of these tropical cyclones travel along the east coast of continents in a poleward direction. The tracks vary from year to year, and vary within any one year (Fig. 16.20). Some of the tracks in the Atlantic might never hit N. America. Other tracks can be along the East Coast of North America, causing significant damage to coastal cities. Yet other tracks continue further westward across the Caribbean Sea and into the Gulf of Mexico before reaching the mainland.

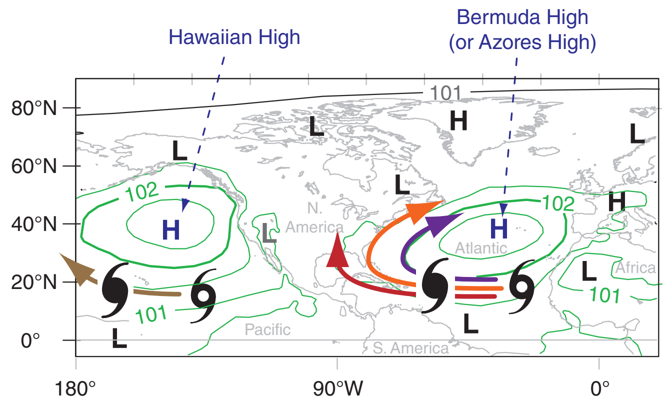


Figure 16.20
Monsoon sea-level pressure (kPa) pattern for July (copied from the General Circulation chapter), with isobars as thin green lines. Winds and average tropical cyclone tracks (dark colored arrows) tend to follow the isobars around the monsoon highs, although actual tracks are quite variable.

16.3.4.2. Extratropical Transition

As the monsoon circulations move the tropical cyclones poleward, the cyclones leave the tropics and subtropics and enter the mid-latitudes. Here, prevailing winds are from the west in the global circulation. Also, winds on the poleward side of the monsoon highs are from the west. Thus, many tropical cyclone tracks eventually turn towards the east at midlatitudes.

Extratropical transition is the name given to the movement of tropical cyclones out of the tropics and into the midlatitudes. At these latitudes the storms are generally moving over colder water, causing them to rapidly die. Nonetheless, these former tropical cyclones can change into extratropical cyclones and still cause damaging floods, wind, and waves if over coastal areas. Even dying tropical cyclones well away from a coastline can influence mid-latitude weather downwind by pumping copious amounts of moisture into the air. This humid air can later serve as the fuel for mid-latitude cyclones thousands of kilometers downwind, creating heavy rain and possible flooding.

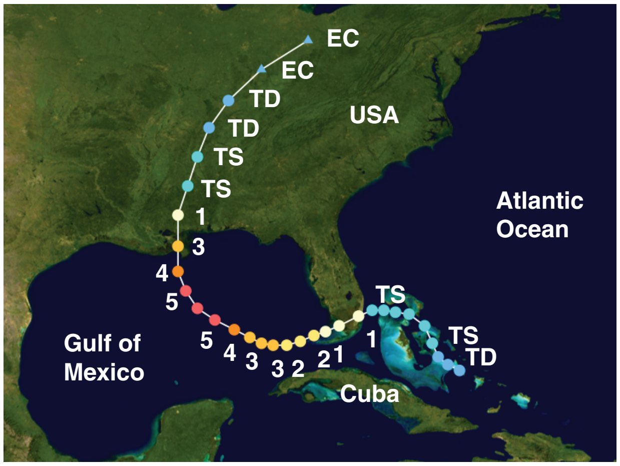


Figure 16.21
Track of hurricane Katrina, showing its intensity during 23 - 31 Aug 2005. 1 to 5 is Saffir-Simpson category. TS = tropical storm. TD = tropical depression. EC = extratropical cyclone.

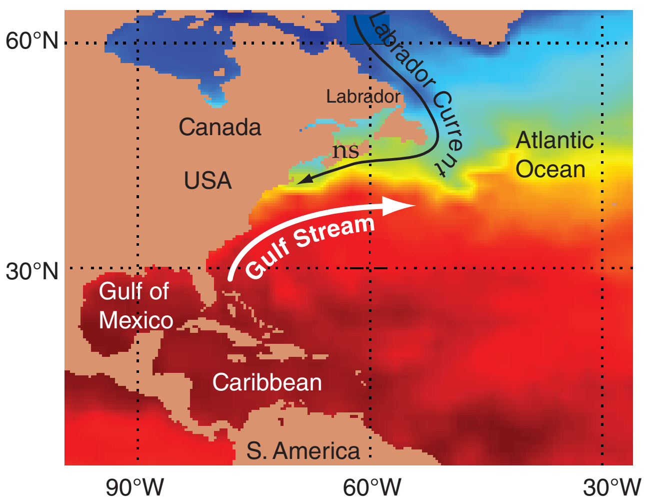
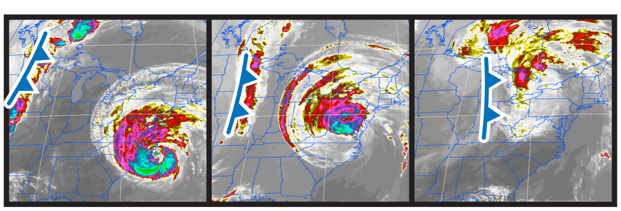


Figure 16.22
Enlargement of part of Fig. 16.8, showing sea-surface temperatures (warm = red fill; cold = blue fill). The Labrador Current (black arrow) and Gulf Stream (white arrow) are shown. ns = Nova Scotia.



(a) 1615 UTC, 18 Sep (b) 0315 UTC, 19 Sep (c) 1615 UTC, 19 Sep

Figure 16.23
Collision of an extratropical cold front with former Hurricane Isabel in Sep 2003. (a) A cold front is over Minnesota, USA, while the hurricane eye makes landfall on the N. Carolina coast. (b) Later. (c) The front and hurricane collide over the Great Lakes. (IR GOES satellite images courtesy of NOAA.)

16.3.5. Tropical Cyclolysis

Tropical cyclones can exist for weeks because they have the ability to create their own fuel supply of warm humid boundary-layer air. They create this fuel by tapping the heat and water stored in the upper ocean. Tropical cyclones die (**tropical cyclolysis**) when they move to a location where they cannot create their own fuel, or if they are destroyed by larger-scale weather systems, as described next.

16.3.5.1. Movement over Land

If the global circulation steers a tropical cyclone over land, it begins to die rapidly because it is unable to tap the warm ocean for more fuel (Fig. 16.21). If the land is a small island or narrow isthmus, then the weakened tropical cyclone can possibly re-intensify when it again moves over warm ocean water.

16.3.5.2. Movement over Cold Water

If the global circulation steers a tropical cyclone over water colder than 26.5°C, then it is unable to create new warm humid air as fast the eyewall thunderstorms consume this fuel. As a result, tropical cyclones quickly weaken over cold water.

This often happens when tropical cyclones are steered poleward toward mid-latitudes, where sea surface temperatures (**SST**) are colder (Fig. 16.8). In the Atlantic, the Gulf Stream of warm ocean water moving north along the east coast of the USA allows tropical cyclones there to survive. However, further north, the very cold Labrador Current (Fig. 16.22) moving south from the Arctic along the east coast of Canada causes most tropical cyclones to diminish below tropical cyclone force before making landfall (see INFO Box on next page).

Tropical cyclolysis can also occur if the layer of warm ocean water is too shallow, such that tropical cyclone-induced ocean waves stir colder deep water to the surface. Also, if a tropical cyclone crosses the path of a previous tropical cyclone, the ocean surface along this previous path is often colder due to the wave stirring.

16.3.5.3. Interaction with Mid-latitude Lows

When tropical cyclones move into mid-latitudes, they can encounter the larger, more powerful mid-latitude cyclones (Fig. 16.23) that were discussed in a previous chapter. Strong mid-latitude cold fronts can inject cold air into the tropical cyclone, causing the tropical cyclone to die. The fronts and associated jet stream aloft often have strong wind shear, and can rip apart the tropical cyclones. Sometimes tropical cyclones will move into the path of a mid-latitude cyclone, allowing the mid-latitude cyclone to capture some of the tropical cyclone's energy and consume the tropical cyclone as the two systems merge.

16.4. DYNAMICS

16.4.1. Origin of Initial Rotation

Define **absolute angular momentum** (*AAM*) as the sum of a relative component of angular momentum associated with tropical-cyclone rotation plus a background component due to the rotation of the Earth:

$$AAM = M_{\tan} \cdot R + 0.5 \cdot f_c \cdot R^2 \qquad \bullet(16.1)$$

where the Coriolis parameter is f_c (≈ 0.00005 s^{-1} at 20° latitude), and the tangential component of velocity at distance R from the eye center is M_{tan}.

As near-surface air converges toward the weak low-pressure of an incipient tropical cyclone, absolute angular momentum is conserved for frictionless flow (i.e., for no drag against the sea surface). Eq. (16.1) indicates that there is nonzero *AAM* even if the incipient tropical cyclone has no rotation yet (i.e., even if $M_{tan.initial} = 0$). Equating the resulting initial *AAM* for air starting at distance R_{init} from the eye center, with the final *AAM* after the air has moved closer to the eye (i.e., smaller R_{final}) yields:

$$M_{\tan} = \frac{f_c}{2} \cdot \left(\frac{R_{init}^2 - R_{final}^2}{R_{final}} \right) \qquad (16.2)$$

For real tropical cyclones you should not neglect frictional drag. Thus, you should anticipate that actual winds will be slower than given by eq. (16.2).

16.4.2. Subsequent Spin-up

As the winds accelerate around a strengthening tropical cyclone, centrifugal force increases. Recall from the Atmospheric Forces & Winds chapter that the equation for **gradient-wind** in cylindrical coordinates (eq. 10.32) is:

$$\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta R} = f_c \cdot M_{\tan} + \frac{M_{\tan}^2}{R} \qquad (16.3)$$

where air density is ρ and the pressure gradient in the radial direction is $\Delta P/\Delta R$. In eq. (16.3) the last term gives the centrifugal force. The gradient wind applies at all radii from the center of the storm, and at all altitudes except near the bottom (in the boundary layer) and near the top (in the anvil or cirrus shield region).

Some researchers find it convenient to neglect Coriolis force closer to the center of the stronger tropical cyclones where the winds are faster. For

INFO • Hurricane Juan Hits Canada

The Canadian East Coast only rarely gets hurricanes — one every 3 to 4 years on average. But it gets many former hurricanes that have weakened to Tropical Storm or lower categories during passage over the cold Labrador current (for Atlantic Canada), or due to passage over land (for Central Canada).

On 29 Sep 2003, Hurricane Juan hit Nova Scotia, Canada, as a category 2 hurricane. Two factors made this possible: (1) warm (but colder than 26.5°C) sea-surface water was relatively close to Nova Scotia; and (2) the hurricane was translating north so quickly that it did not have time to spin down over the narrow cold-water region before making landfall.

This storm began as Tropical Depression #15 about 470 km southeast of Bermuda at noon on 25 Sep. Six hours later it became a Tropical Storm, and by noon 26 Sep it became a Hurricane while 255 km east of Bermuda. During the next 3 days it traveled northward over the Gulf Stream, with intensity of category 2 for most of its journey.

When it reached Canada, it dropped 25 to 40 mm of rain in Nova Scotia. The storm surge was 1.5 m, and the maximum wave height was roughly 20 m. Hundreds of thousands of people lost electrical power due to trees falling on power lines, and 8 people were killed in storm-related accidents.

Before Juan, the most remembered "former hurricane" to strike Canada was Hurricane Hazel. It transitioned into a strong extratropical cyclone before reaching Canada, but caused 81 deaths and severe flooding in southern Ontario in mid October 1954.

Since 1887, there have been two hurricanes of category 3, five of category 2, and 26 of category 1. Info about Canadian hurricanes is available from the **Canadian Hurricane Centre**, Halifax, Nova Scotia.

Sample Application

Suppose that air at a latitude of 12° initially has no rotation as it is drawn toward a low-pressure center 500 km away. When the air reaches 200 km from the low center, what will be its relative tangential velocity?

Find the Answer

Given: R_{init} = 500 km, ϕ = 12°, R_{final} = 200 km
Find: M_{tan} = ? m s^{-1}

Compute the Coriolis parameter:
$f_c = 2 \cdot \omega \cdot \sin(\phi) = (1.458\times10^{-4}\ s^{-1}) \cdot \sin(12°)$
$= 0.0000303\ s^{-1}$

Use eq. (16.2): M_{tan} =

$$\frac{(0.0000303 s^{-1})}{2} \cdot \left(\frac{(500km)^2 - (200km)^2}{200km} \right) = \underline{\mathbf{15.9\ m\ s^{-1}}}$$

Check: Units OK. Physics OK.
Exposition: This is halfway toward the 32 m s^{-1} that defines a Category 1 tropical cyclone. Tropical cyclones cannot exist at the equator because $f_c = 0$ there.

Sample Application

Given a tropical cyclone with: sea-level pressure difference between the eye and far outside of ΔP_{max} = 11.3 kPa, eyewall outside radius of R_o = 25 km, latitude = 14°N, and ρ = 1 kg m^{-3}. From the tropical cyclone model presented later in this chapter, assume the pressure gradient is roughly: $\Delta P/\Delta R = (4/5)\cdot\Delta P_{max}/R_o$.

a) Find the magnitude of the gradient wind just outside of the eye wall (i.e., at $R = R_o$) .

b) Compare the relative importance of the Coriolis and centrifugal terms in the gradient wind eq.

c) Compare this gradient wind with the max expected surface wind of 67 m s^{-1}, and explain any difference.

Find the Answer

Given: ΔP_{max} = 11.3 kPa, R_o = 25 km,
latitude = 14°N, ρ = 1 kg m^{-3}.

Find: M_{tan} = ? m s^{-1}. Compare terms.

a) Use: $\Delta P/\Delta R = (4/5)\cdot\Delta P_{max}/R_o$

Find f_c = (1.458x10^{-4} s^{-1})·sin(14°) = 3.5x10^{-5} s^{-1} .

Use eq. (16.3) and neglect translation speed:

$$\frac{(4/5)}{\rho}\cdot\frac{\Delta P_{\max}}{R_o} = f_c \cdot M_{\tan} + \frac{M_{\tan}^2}{R}$$

Pressure Gradient = Coriolis + Centrifugal

Using trial and error (i.e., trying different values of M_{tan} in a spreadsheet until the left and right sides of the eq balance): M_{tan} = **<u>94.65 m s</u>**$^{-1}$.

b) From my spreadsheet, the terms in the eq. are:

0.361 = 0.003 + 0.358

Pressure Gradient = Coriolis + Centrifugal

Thus, the **<u>Coriolis force is about 2 orders of magnitude smaller than centrifugal force</u>**, at this location of strongest tangential wind just outside the eye wall. At this location, the cyclostrophic assumption (of neglecting Coriolis force) is OK.

c) The max expected <u>surface</u> wind of 67 m s^{-1} includes the effect of **<u>drag against the sea surface</u>**, which is why it is smaller than our answer from (a) M_{tan} = 95 m s^{-1}.

Check: Units OK. Physics OK.
Exposition: This scenario is similar to the composite tropical cyclone model presented later in this chapter.

Sample Application

Suppose a tropical cyclone has a pressure gradient of 1 kPa/15 km at radius 60 km. What is the value of the cyclostrophic wind? Given ρ = 1 kg m^{-3}.

Find the Answer

Given: R = 60 km, ΔP = 1 kPa, ΔR = 15 km,
Find: M_{cs} = ? m s^{-1}

Use eq. (16.4): $M_{cs} = \sqrt{\dfrac{(60\text{km})}{(1\text{kg}\cdot\text{m}^{-3})}\cdot\dfrac{(1000\text{Pa})}{(15\text{km})}}$ = **<u>63 m s</u>**$^{-1}$

Check: Physics & Units OK.
Exposition: This is a Category 4 tropical cyclone.

this situation, the tangential tropical cyclone winds can be crudely approximated by the **cyclostrophic wind**, M_{cs} (see the Atmospheric Forces & Winds Chapter):

$$M_{cs} = M_{\tan} = \sqrt{\frac{R}{\rho}\cdot\frac{\Delta P}{\Delta R}} \tag{16.4}$$

Nonetheless, the gradient-wind equation (16.3) is the most appropriate equation to use for tropical cyclones, at <u>middle</u> altitudes of the storm.

16.4.3. Inflow and Outflow

At the <u>bottom</u> of the tropical cyclone (Fig. 16.24), drag against the sea surface causes the winds to spiral in towards the eyewall. The **boundary-layer gradient wind** equation (see the Atmospheric Forces & Winds chapter) describes this flow well. Without this drag-related inflow, the eyewall thunderstorms would not get sufficient warm humid air to survive, causing the tropical cyclone to die.

Explaining the outflow at the <u>top</u> of a tropical cyclone is trickier, because drag forces are so small that we cannot invoke the boundary-layer gradient-wind equation. One important process is the rapid upward movement of air by thunderstorm updrafts in the eyewall, which deposits enough air molecules at

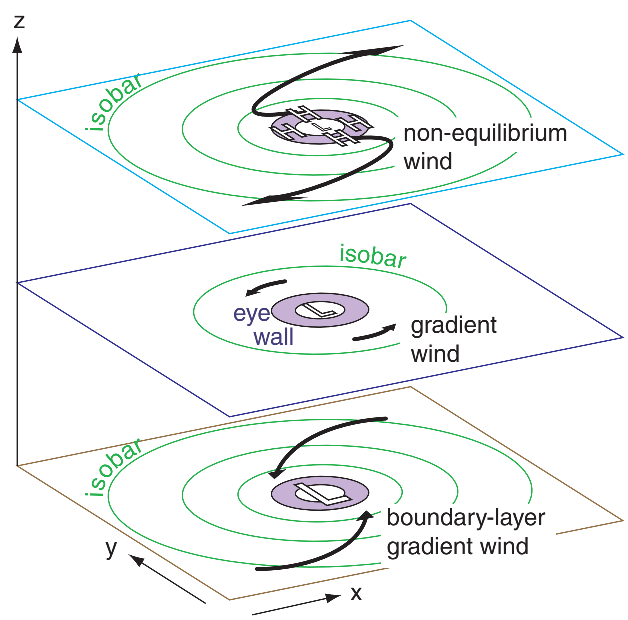


Figure 16.24
Wind dynamics at the bottom, middle and top of a tropical cyclone. Thick arrows represent wind vectors. Low pressure 𝕃 at storm bottom is associated with spiraling inflow winds. High pressure ℍ at storm top is associated with outflow winds. Thus, storm top is the exhaust system of the tropical cyclone engine, and storm bottom is the intake system.

the top of the storm to contribute to high pressure there. Thus, the outflow is related to two factors:

(1) The cyclonically-moving air from the boundary layer is brought to the top of the tropical cyclone by eyewall-thunderstorm updrafts so quickly that its inertia prevents it from instantly changing to anticyclonic outflow. Namely, the outflow is initially moving the wrong way (cyclonically) around the high (Fig. 16.24). The outflow must change direction and increase its speed, and thus is not in steady state.

(2) It is physically impossible to create a balanced gradient-wind flow around a high-pressure area that is surrounded by an excessively strong pressure gradient, as was explained in the Atmospheric Forces & Winds chapter (also see HIGHER MATH in the next column). But the thunderstorm updrafts help create such an excessive high pressure at storm top that the pressure-gradient force exceeds the compensating Coriolis force (Fig. 16.25). The result is a net outward force that causes the air to accelerate outward from the eyewall as a non-equilibrium wind.

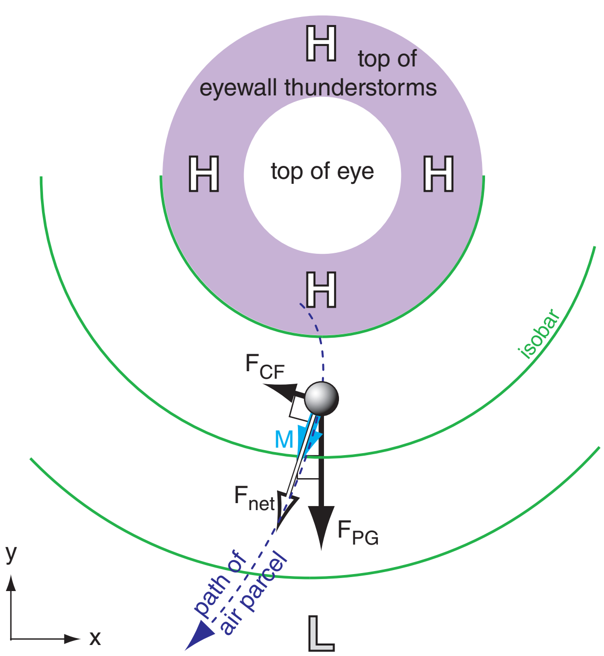


Figure 16.25
Isobars (green lines) on a constant height surface near the top of a tropical cyclone. Purple ring represents the eyewall thunderstorms. H and L are high and low pressure. M is non-equilibrium wind speed (cyan colored). F_{CF} is Coriolis force, F_{PG} is pressure-gradient force, and F_{net} is the vector sum of forces.

HIGHER MATH • Non-equilibrium winds at the top of tropical cyclones

Recall from the Atmospheric Forces & Winds chapter that solutions for the gradient wind around an anticyclone are physically realistic only for curvature Rossby numbers $Ro_c \le 1/4$. But $Ro_c = (\partial P/\partial R)/(\rho \cdot f_c^{\,2} \cdot R)$. Setting $Ro_c = 1/4$, separating variables, and integrating from $P = P_o$ at $R = 0$, to P at R gives:

$$P = P_o - a \cdot R^2 \qquad \text{(a)}$$

where $a = \rho \cdot f_c^{\,2}/8$. This equation gives the max change of pressure with radial distance R that is allowed for a gradient wind that is in equilibrium. But real tropical cyclones can have greater-than-equilibrium pressure gradients at the storm top.

For example, at tropical cyclone top ($z \approx 17$ km) suppose P_o = 8.8kPa in the eye and ρ = 0.14 kg m^{-3}. Thus, $a \approx$ 2.5x10^{-8} kPa km^{-1} at 15° latitude. The strongest horizontal pressure gradient possible for a gradient wind around the high-pressure center at the top of the eye is plotted below as the slope of the solid line (from equation a). This is a minuscule pressure gradient, due in part to the very small Coriolis force at 15° latitude.

But actual horizontal pressure gradients can be much stronger (dashed line). Thus, steady-state gradient winds (winds that follow the curved isobars) are not possible at tropical-cyclone top. Instead, the non-equilibrium winds are accelerating outward (crossing this isobars at a large angle) from the high due to the strong pressure gradient, and only gradually develop some anticyclonic rotation further from the eye.

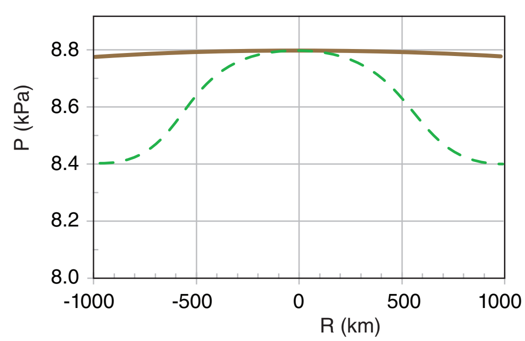


Fig. 16.b. *Horizontal pressure gradients at the top of a tropical cyclone. R = distance from cyclone center, P = atmospheric pressure. The slope of the solid brown line shows the max pressure gradient that admits a gradient-wind solution. The dashed green line is hypothetical and focuses on the pressure gradient from the eyewall outward.*

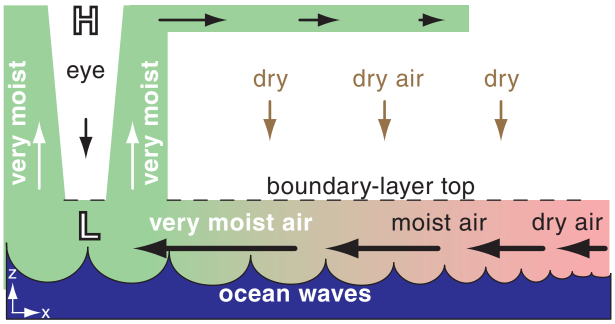


Figure 16.26
The intake system in the bottom of a tropical cyclone. As boundary-layer winds accelerate, they cause larger ocean waves that add more moisture (light green indicates high humidity) to the inflowing air (thick arrows). L is the low-pressure center at sea level. Air is exhausted out of the top of the storm in the central dense overcast composed of thunderstorm anvils.

INFO • Tropical Cyclone Condensation Energy

One way to estimate the energy of tropical cyclones is by the amount of latent heat released during condensation. Rainfall rate is a measure of the net condensation.

An average Atlantic Hurricane produces about 1.5 cm of rain day^{-1}, which when accumulated over the area (of radius 665 km) covered by hurricane rain is equivalent to **2.1x10**^{10} m^3 day^{-1}.

This converts to **6x10**^{14} W (= 5.2x10^{19} J day^{-1}). Namely, it is about 200 times the daily electricity generation capacity of the whole world.

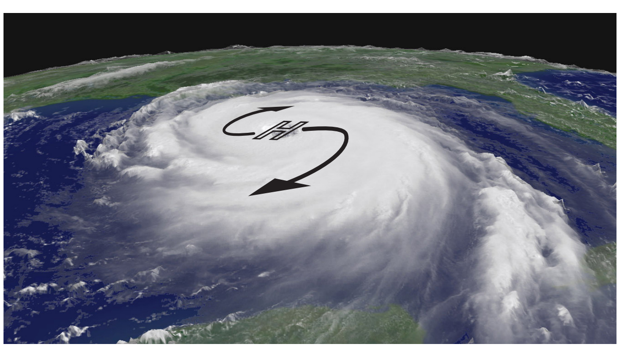


Figure 16.27
Exhaust system at the top of a tropical cyclone, where H is the high-pressure center near the tropopause and arrows are outflow winds. Computer-enhanced 3-D satellite image of Hurricane Katrina over the Gulf of Mexico, taken 28 Aug 2005 at 1545 UTC. Visible satellite imagery is mapped into a 3-D surface of cloud-top altitudes estimated from satellite IR data. (Image courtesy of US DOC/NOAA.)

16.5. THERMODYNAMICS

Tropical cyclones work somewhat like engines. There is an intake system (the atmospheric boundary layer) that draws in the fuel (warm, humid air). The engine (thunderstorms) converts heat into mechanical energy (winds and waves). And there is an exhaust system (precipitation fallout for water and anvil blowout for air) for the spent fuel.

16.5.1. Fuel Creation and Intake

In midlatitudes, thunderstorms can last for hours. However, tropical cyclones (which are made of thunderstorms) can last for weeks. Why the difference in longevity?

The main reason is that mid-latitude thunderstorms use warm humid air from only the nearby boundary layer, and after this nearby fuel is consumed the thunderstorms die. Even supercells and squall-line storms rely on an existing fuel supply of warm humid air, which they can utilize as they move across the countryside.

However, tropical cyclones create their own fuel of warm humid boundary-layer air. They do this via the fast near-surface winds that create large ocean waves (Fig. 16.26). These violent waves break and foam, causing rapid evaporation of sea-spray water into the air, and efficient transfer of heat from the ocean surface to the air. By the time this air reaches the base of the eyewall, it has an air temperature nearly equal to the sea-surface temperature and a relative humidity of nearly 100%.

Namely, tropical cyclones extract heat from the ocean. The ocean is a giant heat reservoir that has been absorbing sunlight all Summer and early Fall. This is the reason why warm, deep sea-surface temperatures are needed for tropical cyclones.

16.5.2. Exhaust

The large volume of inflowing boundary-layer air is good and bad for the tropical cyclone. It is good because this air carries the sensible and latent heat fuel for the storm. It is bad because it also brings massive amounts of other air molecules (nitrogen, oxygen) into the core of the storm.

The problem is that if these air molecules were to accumulate in the center of the storm, their weight would cause the sea-level air pressure to rise. This would weaken the surface low pressure, which would reduce the radial pressure gradient near sea level, causing the inflow winds and waves to diminish. With reducing inflow, the eyewall thunderstorms would run out of fuel and die, causing the tropical cyclone to disintegrate in half a day or so.

However, we know that tropical cyclones can exist for weeks. Thus, there must be some other mechanism that removes air from the tropical cyclone core as fast as it enters. That other mechanism is the strong outflow winds in the thunderstorm anvils at the top of the eyewall. This outflow exists because of high pressure at the top of the troposphere in the eye and eyewall regions (Figs. 16.26 & 16.27), which drives winds outward down the pressure gradient.

But what causes this high pressure aloft? We already discussed a non-hydrostatic process — the air deposited aloft from eyewall thunderstorm updrafts. But another process is hydrostatic — related to the excessive warmth of the center or **core** portion of the tropical cyclone.

16.5.3. Warm Core

Air rises moist adiabatically in the eyewall thunderstorms, releasing a lot of latent heat along the way. But then after losing water due to precipitation from the eyewall, some of the air warms as it descends dry adiabatically in the eye. Thus, the **core** (both the eye and the eye wall) of the tropical cyclone has much warmer temperature than the ambient air outside the storm.

Typical temperature excesses of the **warm core** are 0 to 4°C near sea level, and 10 to 16°C warmer at 12 to 16 km altitude (Fig. 16.28). An INFO box illustrates such a warm-core system.

This radial temperature gradient (warm core vs. cool exterior) causes a radial pressure-gradient reversal with increasing altitude (Fig. 16.29), as is explained by the thermal-wind expression. To determine this pressure gradient, define the sea-level pressure in the eye as $P_{B\ eye}$, and that of the distant surroundings at sea level as $P_{B\ \infty}$. Near the top of the tropical cyclone, define the core pressure as $P_{T\ eye}$ and that of its distant surroundings as $P_{T\ \infty}$ at the same altitude. Subscripts B and T denote bottom and top of the troposphere.

Suppose a tropical cyclone is approximately $z_{max} \approx$ 15 km deep, has a temperature in the eye averaged over the whole tropical cyclone depth of $\overline{T_{eye}}$ = 273 K, and has ambient surface pressure distant from the storm of $P_{B\ \infty}$ = 101.3 kPa. The pressure difference at the top of the tropical cyclone ($\Delta P_T = P_{T\ \infty} - P_{T\ eye}$) is approximately related to the pressure difference at the bottom ($\Delta P_B = P_{B\ \infty} - P_{B\ eye}$) by:

$$\Delta P_T \approx a \cdot \Delta P_B - b \cdot \Delta T \qquad \bullet(16.5)$$

where $a \approx 0.15$ (dimensionless), $b \approx 0.7$ kPa K^{-1}, and $\Delta T = T_{eye} - T_{\infty}$ is the temperature difference averaged over the troposphere. Note that ΔP_T is negative.

Next, consider the tangential winds. These winds spiral cyclonically around the eye at low al-

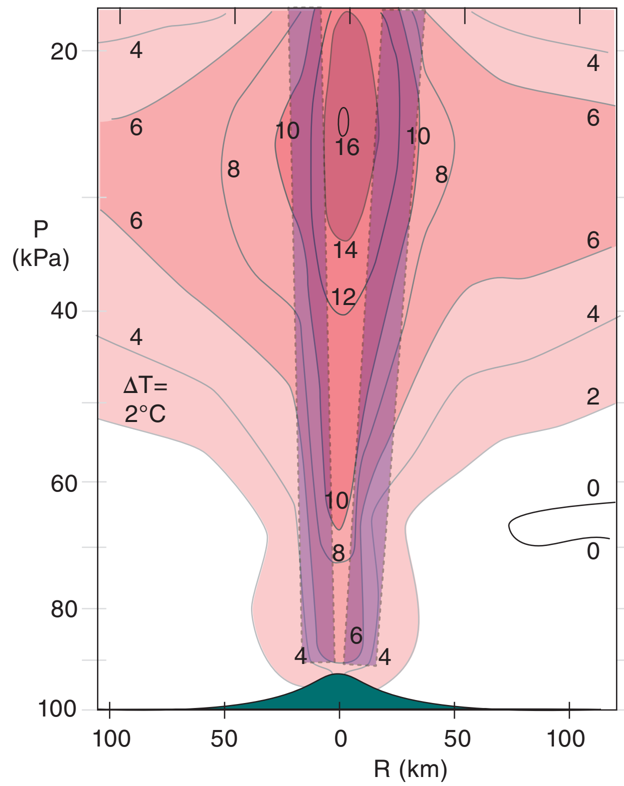


Figure 16.28
The exceptionally warm core of Hurricane Inez. Vertical cross section through the eye, showing the temperature excess ΔT compared to the environmental temperature at the same altitude. The approximate eyewall location is highlighted in purple. R is radial distance from the center of the eye, and P is pressure. Inez existed during 22 Sep to 11 Oct 1996 in the eastern Caribbean, and reached category 4 intensity. The green area at the bottom of the figure masks atmospheric pressures that did not exist at sea level, because of the very low surface pressure in and near the eye. (Based on data from Hawkins and Imbembo, 1976.)

Sample Application

The surface pressure in the eye of a tropical cyclone is 90 kPa, while the surrounding pressure is 100 kPa. If the core is 10 K warmer than the surroundings, find the pressure difference at the top of the tropical cyclone between the eye and surroundings.

Find the Answer

Given: ΔP_B =100 – 90 = 10 kPa, ΔT = 10 K
Find: ΔP_T = ? kPa

Use eq. (16.5):
ΔP_T = (0.15)·(10kPa) – (0.7kPa K^{-1})·(10K) = **– 5.5 kPa**

Check: Units OK. Physics OK.
Exposition: This answer is negative, meaning that the eye has higher pressure than the surroundings, aloft. This pressure reversal drives the outflow aloft.

INFO • Warm vs. Cold Core Cyclones

For a cyclone to survive and intensify, air must be constantly withdrawn from the top. This removal of air mass from the cyclone center (**core**) counteracts the inflow of boundary-layer air at the bottom, which always happens due to surface drag. The net result of storm-top outflow is storm-bottom low pressure in the core, which drives the near-surface winds.

Warm-core cyclones (e.g., tropical cyclones) and **cold-core cyclones** (e.g., extratropical lows) differ in the way they cause horizontal divergence to remove air from the top of the cyclones [see Fig. 16.c, parts (a) and (b)].

Tropical cyclones are vertically stacked, with the storm-top eye of the tropical cyclone almost directly above the storm-bottom eye [Fig. 16.c(a)]. Intense latent heating in the tropical cyclone warms the whole depth of the troposphere near the core, causing high pressure aloft because warm layers of air have greater thickness than cold. This high pressure aloft causes air to diverge horizontally at the top of the tropical cyclone, which is why visible and IR satellite loops show cirrus and other high clouds flowing away from the tropical cyclone center.

Extratropical lows are not vertically stacked, but have low pressure that tilts westward with increasing height [Fig 16.c(b)]. As the surface circulation around the cyclone advects in cold, polar, boundary-layer air on the west side of the cyclone, the small thicknesses in that sector cause pressure to decrease more rapidly with height. The net result is an upper-level trough west of, and a ridge east of, the surface low. A jet stream meandering through this trough-ridge system would cause horizontal divergence (D), as was shown in the Extratropical Cyclone chapter.

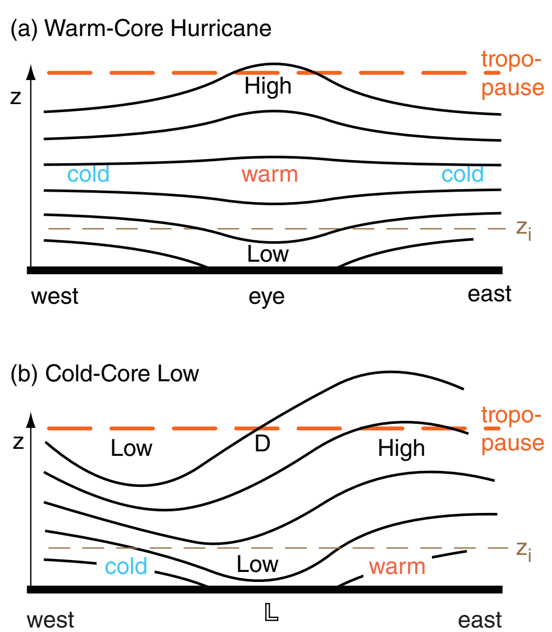


Figure 16.c.
Vertical cross section of (a) warm, and (b) cold core cyclones. Thin lines are isobars, and z_i is ABL top.

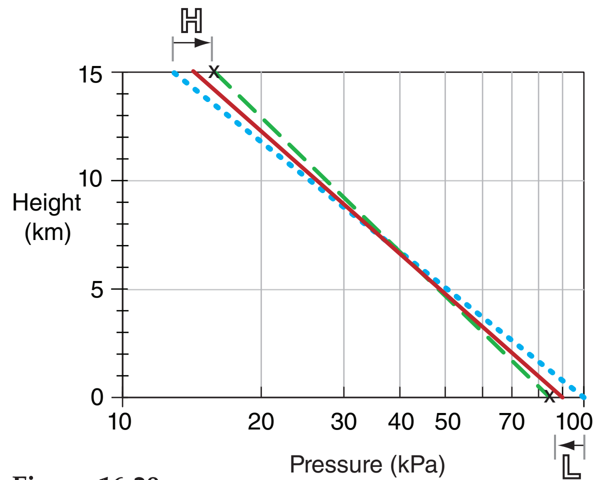


Figure 16.29
Blue dotted line: hydrostatic pressure decrease for cool air outside the tropical cyclone. Red solid line: hydrostatic vertical pressure gradient in the warm core. The core has higher pressure relative to its environment near the top of the storm, even though the core has lower pressure near sea level (z = 0), as explained by the hypsometric eq. Additional nonhydrostatic P variation (green dashed line between two x's) in the vertical is due to strong updrafts in eyewall thunderstorms, removing low-altitude air and depositing it at storm top. This can cause non-hydrostatic descent of warm air in the eye. (This is a semi-log graph.)

titudes, but spiral anticyclonically near the top of the troposphere well away from the eye. Hence, the tangential velocity must decrease with altitude and eventually change sign. The ideal gas law can be used with the gradient wind eq. (16.3) to show how tangential wind component M_{tan} varies with altitude z :

$$\left(\frac{2 \cdot M_{\tan}}{R} + f_c\right) \cdot \frac{\Delta M_{\tan}}{\Delta z} = \frac{|g|}{\overline{T}} \cdot \frac{\Delta T}{\Delta R} \qquad \bullet(16.6)$$

Sample Application

Suppose at a radius of 40 km the tangential velocity decreases from 50 m s^{-1} at the surface to 10 m s^{-1} at 10 km altitude. Find the radial temperature gradient at a latitude where the Coriolis parameter is 0.00005 s^{-1}. Assume $|g|/T = 0.0333$ m·s^{-2}·K^{-1} .

Find the Answer

Given: $R = 40$ km, $M_{tan} = 50$ m s^{-1} at $z = 0$,
$M_{tan} = 10$ m s^{-1} at $z = 10$ km, $f_c = 0.00005$ s^{-1} ,
$|g|/T = 0.0333$ m·s^{-2}·K^{-1}

Find: $\Delta T/\Delta R$ = ? K km^{-1}

Use eq. (16.6):

$$\frac{|g|}{\overline{T}} \cdot \frac{\Delta T}{\Delta R} = \left(\frac{2 \cdot (30\text{m/s})}{40,000\text{m}} + 0.00005\text{s}^{-1}\right) \cdot \frac{(10 - 50\text{m/s})}{(10,000\text{m})}$$

$$= -6.2\text{x}10^{-6}\ \text{s}^{-2}.$$

Thus, $$\frac{\Delta T}{\Delta R} = \frac{-6.2 \times 10^{-6}\text{s}^{-2}}{0.0333\text{m} \cdot \text{s}^{-2} \cdot \text{K}^{-1}} = \underline{\mathbf{-0.19\ K\ km^{-1}}}.$$

Check: Units OK. Physics OK.

Exposition: From the center of the eye, the temperature decreases about 7.4 K at a radius of 40 km, for this example. Indeed, the core is warm.

HIGHER MATH • Pressure Reversal

Derivation of eq. (16.5):

Use the hypsometric equation to relate the pressure at the top of the tropical cyclone eye to the pressure at the bottom of the eye. Do the same for the surroundings. Use those equations to find the pressure difference at the top:

$$\Delta P_T = P_{B\infty}\cdot \exp[-|g|\cdot z_{max} / (\Re\cdot \overline{T}_{\infty})] - P_{B.eye}\cdot \exp[-|g|\cdot z_{max} / (\Re\cdot \overline{T}_{eye})]$$

Then, using $P_{B.eye} = P_{B\infty} - \Delta P_B$, collecting the exponential terms that are multiplied by $P_{B\infty}$, and finally using a first-order series approximation for those exponentials, one gets eq. (16.5), where

$a = \exp[-|g|\cdot z_{max} / (\Re\cdot \overline{T}_{eye})] \approx 0.15$,
$b = -(|g|\cdot z_{max}\cdot P_{B\infty}) / (\Re\cdot \overline{T}_{eye}\cdot \overline{T}_{\infty}) \approx 0.7$ kPa K^{-1},
$|g|$ = 9.8 m·s^{-2} is gravitational acceleration magnitude, and
$\Re$ = 287.04 m^2·s^{-2}·K^{-1} is the gas constant for dry air.

HIGHER MATH • Warm Core Winds

Derivation of eq. (16.6):

There are 3 steps: (1) scale analysis; (2) differentiation of eq. (16.3) with respect to height z; and (3) simplification of the pressure term.

(1) Scale Analysis

Differentiate the ideal gas law $P = \rho\cdot\Re\cdot T$ with respect to z, and use the chain rule of calculus:

$$\frac{\partial P}{\partial z} = \rho\Re\frac{\partial T}{\partial z} + \Re T\frac{\partial \rho}{\partial z}$$

Then divide this eq. by P use the ideal gas law on the right side:

$$\frac{1}{P}\frac{\partial P}{\partial z} = \frac{1}{T}\frac{\partial T}{\partial z} + \frac{1}{\rho}\frac{\partial \rho}{\partial z}$$

Between z = 0 to 20 km, typical variations of the variables are: P = 101 to 5.5 kPa, T = 288 to 216 K, and ρ = 1.23 to 0.088 kg m^{-3} . Thus, the temperature term in the eq. above varies only 1/4 as much as the other two terms. Based on this scale analysis, we can neglect the temperature term, which leaves:

$$\frac{1}{P}\frac{\partial P}{\partial z} \approx \frac{1}{\rho}\frac{\partial \rho}{\partial z} \qquad \text{(a)}$$

(2) Differentiate eq. (16.3) with respect to z:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = f_c\cdot\frac{\partial M_{tan}}{\partial z} + \frac{2M_{tan}}{R}\cdot\frac{\partial M_{tan}}{\partial z}$$

Upon switching the left and right sides:

$$\left[\frac{2M_{tan}}{R} + f_c\right]\cdot\frac{\partial M_{tan}}{\partial z} = \frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] \qquad \text{(b)}$$

(3) Simplify the pressure term:

First, use the chain rule:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = \frac{1}{\rho}\cdot\frac{\partial}{\partial z}\left[\frac{\partial P}{\partial R}\right] + \frac{\partial P}{\partial R}\cdot\frac{\partial}{\partial z}\left[\rho^{-1}\right]$$

R and z are independent, allowing the order of differentiation to be reversed in the first term on the RHS:

(continues in next column)

HIGHER MATH • Warm Core Winds 2

Derivation of eq. (16.6) *(continuation)*

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = \frac{1}{\rho}\cdot\frac{\partial}{\partial R}\left[\frac{\partial P}{\partial z}\right] - \frac{1}{\rho^2}\frac{\partial P}{\partial R}\cdot\frac{\partial \rho}{\partial z}$$

Use the hydrostatic eq. $\partial P/\partial z = -\rho\cdot|g|$ in the first term on the RHS:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = \frac{1}{\rho}\cdot\frac{\partial}{\partial R}\left[-\rho\cdot|g|\right] - \frac{1}{\rho}\frac{\partial P}{\partial R}\cdot\frac{1}{\rho}\frac{\partial \rho}{\partial z}$$

But $|g|$ is constant. Also, substitute (a) in the last term:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = -\frac{|g|}{\rho}\cdot\frac{\partial \rho}{\partial R} - \frac{1}{\rho}\frac{\partial P}{\partial R}\cdot\frac{1}{P}\frac{\partial P}{\partial z}$$

Substitute the ideal gas law in the first term on the RHS:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = -\frac{|g|}{\rho\cdot\Re}\cdot\frac{\partial(P\cdot T^{-1})}{\partial R} - \frac{1}{P}\frac{\partial P}{\partial R}\cdot\frac{1}{\rho}\frac{\partial P}{\partial z}$$

Use the chain rule on the first term on the right, and substitute the hydrostatic eq. in the last term:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = -\frac{P\cdot|g|}{\rho\cdot\Re}\cdot\frac{\partial(T^{-1})}{\partial R} - \frac{|g|}{\rho\cdot\Re\cdot T}\cdot\frac{\partial P}{\partial R} + \frac{|g|}{P}\frac{\partial P}{\partial R}$$

Substitute the ideal gas law in the 2nd term on the right:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = \frac{P\cdot|g|}{\rho\cdot\Re\cdot T^2}\cdot\frac{\partial T}{\partial R} - \frac{|g|}{P}\cdot\frac{\partial P}{\partial R} + \frac{|g|}{P}\frac{\partial P}{\partial R}$$

But the last two terms cancel. Using the ideal gas law in the remaining term leaves:

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho}\frac{\partial P}{\partial R}\right] = \frac{|g|}{T}\cdot\frac{\partial T}{\partial R} \qquad \text{(c)}$$

(4) Completion:

Finally, equate (b) and (c):

$$\boxed{\left[\frac{2M_{tan}}{R} + f_c\right]\cdot\frac{\partial M_{tan}}{\partial z} = \frac{|g|}{T}\cdot\frac{\partial T}{\partial R}} \qquad (16.6)$$

which is the desired answer, when converted from derivatives to finite differences.

Table 16-5. Example of thermodynamic states within a tropical cyclone, corresponding to the points (Pt) circled in Fig. 16.30. Reference: T_o = 273 K, P_o = 100 kPa.

Pt	P (kPa)	T (°C)	T_d (°C)	r (g/kg)	s [J/(K·kg)]
1	100	28	–70	≈ 0	98
2	90	28	28	28	361
3	25	–18	–18	3.7	366
4	20	–83	–83	≈ 0	98

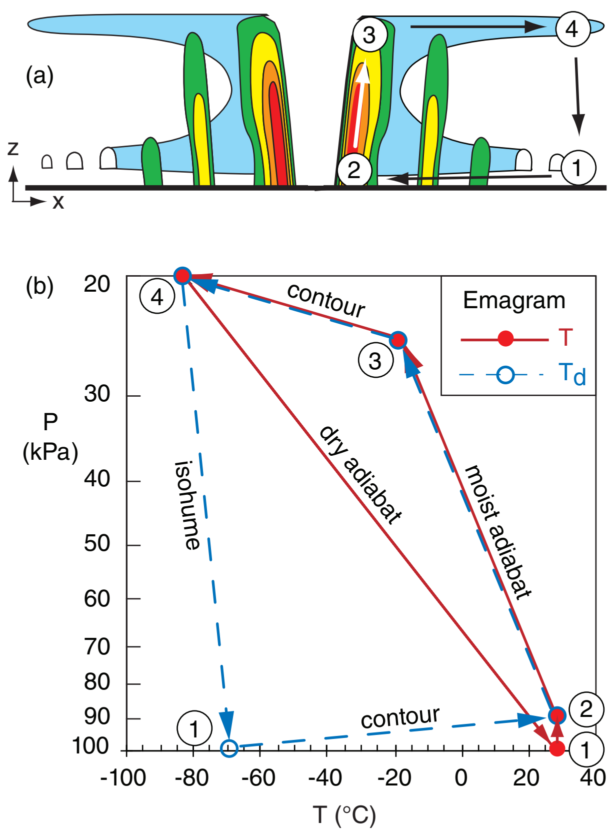


Figure 16.30
(a) Circulation of air through the tropical cyclone. (b) Thermodynamic diagram showing tropical cyclone processes for the same numbered locations as in (a). Solid red lines show the temperature changes, dashed blue lines show dew-point-temperature changes.

where T is absolute temperature, and the overbar denotes an average over depth. Eqs. (16.5 & 16.6 are overly simplistic because they ignore nonhydrostatic and non-equilibrium effects.

16.5.4. Carnot-cycle Heat Engine

Tropical cyclones are analogous to **Carnot heat engines** in that they convert thermal energy into mechanical energy. One measure of the energy involved is the **total entropy** s per unit mass of air:

$$s = C_p \cdot \ln\left(\frac{T}{T_o}\right) + \frac{L_v \cdot r}{T} - \Re \cdot \ln\left(\frac{P}{P_o}\right) \qquad \bullet(16.7)$$

where C_p = 1004 J·kg^{-1}·K^{-1} is the specific heat of air at constant pressure, T is absolute temperature, L_v = 2500 J $g_{water\ vapor}^{-1}$ is the latent heat of vaporization, r is mixing ratio, $\Re$ = 287 J·kg^{-1}·K^{-1} is the gas-law constant, P is pressure. T_o = 273 K and P_o = 100 kPa are arbitrary reference values.

A thermo diagram is used to illustrate tropical cyclone thermodynamics (Fig. 16.30). Because of limitations in range and accuracy of thermo diagrams, you might find slightly different answers than those given in Table 16-5. For simplicity, assume constant sea surface temperature of T_{SST} = 28°C.

As an initial condition at Point 1 in Fig. 16.30, consider relatively warm (T = 28°C) dry air (r ≈ 0, T_d = –70°C) in the boundary layer (z = 0; P = 100 kPa), but outside of the tropical cyclone. Using eq. (16.7), the initial entropy is s_1 ≈ 98 J·kg^{-1}·K^{-1}.

In the first part of the tropical cyclone's Carnot cycle, air in the boundary layer spirals in from Point 1 toward the eye wall of the tropical cyclone (Point 2) isothermally (T = 28°C, because of heat transfer with the sea surface) at constant height (z = 0 = sea level). Pressure decreases to P = 90 kPa as the air approaches the low-pressure eye. Evaporation from the sea surface increases the mixing ratio to saturation (r ≈ 28 g kg^{-1}, T_d ≈ 28°C), thereby causing entropy to increase to s_2 ≈ 361 J·kg^{-1}·K^{-1}. This evaporation is the major source of energy for the storm.

From Point 2, air rises moist adiabatically to Point 3 in the thunderstorms of the eye wall. The moist-adiabatic process conserves entropy (s_3 = 366 J·kg^{-1}·K^{-1} , within the accuracy of the thermo diagram, thus $s_2 \approx s_3$). During this process, temperature drops to T ≈ –18°C and mixing ratio decreases to about r ≈ 3.7 g kg^{-1}. However, the decrease of pressure (P = 90 to 25 kPa) in this rising air compensates to maintain nearly constant entropy.

Once the cloudy air reaches the top of the troposphere at Point 3, it flows outward to Point 4 at roughly constant altitude. (Note that, the thermo diagrams at the end of the Atmospheric Stability chapter do not go high enough to simulate a real tropical cyclone of 15 km depth, so we will use $z \approx$ 10 km here.) The divergence of air is driven by a pressure gradient of P = 25 kPa in the eye to 20 kPa outside the tropical cyclone.

During this high-altitude outflow from Points 3 to 4, air rapidly loses heat due to infrared radiation, causing its temperature to decrease from $T = T_d$ = –18°C to – 83°C. The air remains saturated, and mixing ratio decreases from $r \approx 3.7$ to near 0. The cooling also converts more water vapor into precipitation. Entropy drops to 98 J·kg^{-1}·K^{-1}.

Finally, the air subsides dry adiabatically from Points 4 to 1, with no change of mixing ratio ($r \approx 0$; $T_d \approx$ –70°C). Temperature increases adiabatically to T = 28°C, due to compression as the air descends into higher pressure (P = 100 kPa). This dry adiabatic process also preserves entropy, and thus is called an **isentropic** process (and dry adiabats are also known as **isentropes**). The final state of the air is identical to the initial state, at Point 1 in Fig. 16.30.

This Carnot process is a closed cycle – the air can recirculate through the tropical cyclone. However, during this cycle, entropy is gained near the sea surface where the temperature is warm, while it is lost near cloud top where temperatures are colder.

The gain of entropy at one temperature and loss at a different temperature allows the Carnot engine to produce mechanical energy ME according to

$$ME = (T_B - T_{T.avg})\cdot(s_{eyewall} - s_\infty)_B \qquad \bullet(16.8)$$

where subscripts B and T denote bottom and top of the troposphere, eyewall denotes boundary-layer air under the eye wall, and ∞ denotes the ambient conditions at large distances from the tropical cyclone (e.g., $P_\infty \approx$ 101.3 kPa). This mechanical energy drives the tropical cyclone-force winds, ocean waves, atmospheric waves, and causes mixing of both the atmosphere and ocean against buoyant forces.

If all of the mechanical energy were consumed trying to maintain the tropical cyclone-force winds against the frictional drag in the boundary layer (an unrealistic assumption), then the tropical cyclone could support the following maximum pressure ratio at the surface:

$$\ln\left(\frac{P_\infty}{P_{eye}}\right)_B = \frac{ME}{T_B \cdot \Re} \qquad (16.9)$$

Sample Application

Suppose air in the eye of a tropical cyclone has the following thermodynamic state: P = 70 kPa, r = 1 g kg^{-1}, T = 15°C. Find the entropy.

Find the Answer

Given: P = 70 kPa, r = 1 g kg^{-1}, T = 288 K.
Find: s = ? J·kg^{-1}·K^{-1} .

Use eq. (16.7):
$$s = \left(1004\frac{J}{kg_{air}\cdot K}\right)\cdot\ln\left(\frac{288K}{273K}\right) + \left(2500\frac{J}{g_{water}}\right)\cdot\left(1\frac{g_{water}}{kg_{air}}\right)\frac{1}{288K} - \left(287\frac{J}{kg_{air}\cdot K}\right)\cdot\ln\left(\frac{70kPa}{100kPa}\right)$$

s = **165 J·kg**$^{-1}$·K^{-1}

Check: Units OK. Physics OK.
Exposition: The actual value of entropy is meaningless, because of the arbitrary constants T_o and P_o. However, the difference between two entropies is meaningful, because the arbitrary constants cancel out.

Sample Application

Find the mechanical energy and minimum possible eye pressure that can be supported by the tropical cyclone of Table 16-5.

Find the Answer

Given: $s_{eyewall}$ = 361 J·kg^{-1}·K^{-1}, s_∞ = 98 J·kg^{-1}·K^{-1}
T_B = 28°C, $T_{T\,avg} \approx$ 0.5·(–18–83) = –50.5°C.
Find: ME = ? kJ kg^{-1}, P_{eye} = ? kPa

Use eq. (16.8): ME = (28 + 50.5)(K)·(361–98)(J·kg^{-1}·K^{-1})
= **20.6 kJ kg**$^{-1}$

Use eq. (16.9):
$$\ln\left(\frac{P_\infty}{P_{eye}}\right)_B = \frac{(20,600\text{J}\cdot\text{kg}^{-1})}{(301\text{K})\cdot(287\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1})} = 0.238$$

Solving for P_{eye} gives:
P_{eye} =(101.3kPa)/exp(0.238) = **79.8 kPa**.

Check: Units OK. Physics OK.
Exposition. This eye pressure is lower than the actual eye pressure of 90 kPa. The difference is related to the ME of winds and waves.

INFO • The Power of Tropical Cyclones

Another measure of tropical cyclone power is the rate of dissipation of kinetic energy by wind drag against the sea surface. This is proportional to the ABL wind speed cubed, times the sea-surface area over which that speed is valid, summed over all areas under the tropical cyclone. K. A. Emanuel (1999, *Weather*, **54**, 107-108) estimates dissipation rates of 3 x 10^{12} W for a Category 3 Atlantic hurricane (with max winds of 50 m s^{-1} at radius 30 km), and 3 x 10^{13} W for a Pacific typhoon (with max winds of 80 m s^{-1} at radius 50 km).

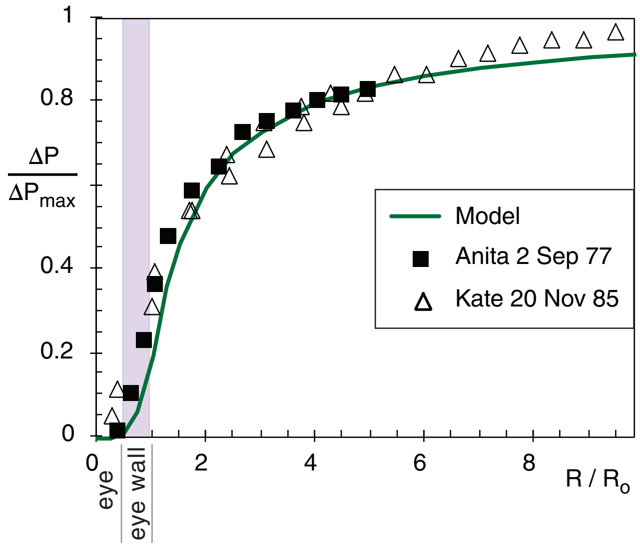


Figure 16.31
Surface pressure distribution across a tropical cyclone. Model is eq. (16.11).

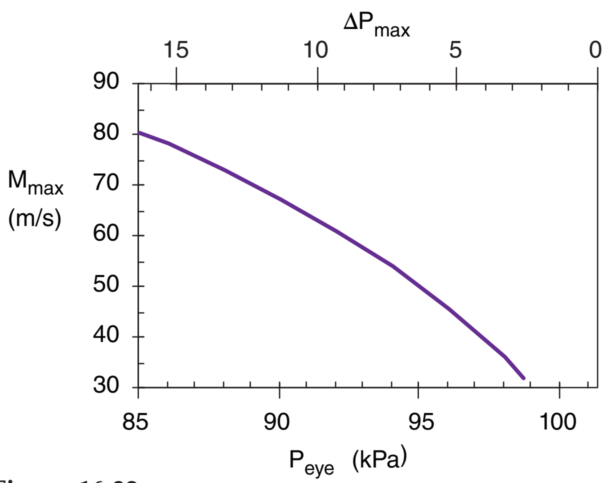


Figure 16.32
Maximum sustained winds around a tropical cyclone increase as sea-level pressure in the eye decreases.

Sample Application

What max winds are expected if the tropical cyclone eye has surface pressure of 95 kPa?

Find the Answer

Given: $P_{B\ eye}$ = 95 kPa, Assume $P_{B\ \infty}$ = 101.3 kPa
Find: M_{max} = ? m s^{-1}.
Assume translation speed can be neglected.

Use eq. (16.12):
M_{max} = [20 (m s^{-1})·kPa$^{-1/2}$]·(101.3–95kPa)$^{1/2}$ = **<u>50 m s^{-1}</u>**

Check: Units OK. Physics OK.
Exposition: This is a category 3 tropical cyclone on the Saffir-Simpson tropical cyclone Wind Scale.

16.6. A TROPICAL CYCLONE MODEL

Although tropical cyclones are quite complex and not fully understood, we can build an idealized model that mimics some of the real features.

16.6.1. Pressure Distribution

Eye pressures of 95 to 99 kPa at sea level are common in tropical cyclones, although a pressure as low as P_{eye} = 87 kPa has been measured. One measure of tropical cyclone strength is the pressure difference ΔP_{max} between the eye and the surrounding ambient environment where P_∞ = 101.3 kPa:

$$\Delta P_{\max} = P_\infty - P_{eye} \tag{16.10}$$

The surface pressure distribution across a tropical cyclone can be approximated by:

$$\frac{\Delta P}{\Delta P_{\max}} = \begin{cases} \dfrac{1}{5}\cdot\left(\dfrac{R}{R_o}\right)^4 & \text{for } R \le R_o \\ \left[1 - \dfrac{4}{5}\cdot\dfrac{R_o}{R}\right] & \text{for } R > R_o \end{cases} \tag{16.11}$$

where $\Delta P = P(R) - P_{eye}$, and R is the radial distance from the center of the eye. This pressure distribution is plotted in Fig 16.31, with data points from two tropical cyclones.

R_o is the critical radius where the maximum tangential winds are found. In the tropical cyclone model presented here, R_o is twice the radius of the eye. Eyes range from 4 to 100 km in radius, with average values of 15 to 30 km. Thus, one anticipates average values of critical radius of $30 < R_o < 60$ km, with an observed range of $8 < R_o < 200$ km.

16.6.2. Tangential Velocity

To be classified as a tropical cyclone, the sustained winds (averaged over 1-minute) must be 33 m s^{-1} or greater near the surface. While most anemometers are unreliable at extreme wind speeds, maximum tropical cyclone winds have been reported in the 75 to 95 m s^{-1} range.

As sea-level pressure in the eye decreases, maximum tangential surface winds M_{max} around the eye wall increase (Fig. 16.32). An empirical approximation for this relationship, based on **Bernoulli's equation** (see the Regional Winds chapter), is :

$$M_{\max} = a\cdot(\Delta P_{\max})^{1/2} \qquad \bullet(16.12)$$

where a = 20 (m s^{-1})·kPa$^{-1/2}$.

If winds are assumed to be cyclostrophic (not the best assumption, because drag against the sea surface and Coriolis force are neglected), then the previous approximation for pressure distribution (eq. 16.11) can be used to give a distribution of tangential velocity M_{tan} (relative to the eye) in the boundary layer:

$$\frac{M_{\tan}}{M_{\max}} = \begin{cases} (R / R_o)^2 & \text{for } R \leq R_o \\ (R_o / R)^{1/2} & \text{for } R > R_o \end{cases} \qquad (16.13)$$

where the maximum velocity occurs at critical radius R_o (assumed to be twice the radius of the eye). This is plotted in Fig. 16.33, with data points from a few tropical cyclones.

For the tropical cyclones plotted in Fig. 16.33, the critical radius of maximum velocity was in the range of R_o = 20 to 30 km. This is a rough definition of the outside edge of the **eye wall** for these tropical cyclones, within which the heaviest precipitation falls. The maximum velocity for these storms was M_{max} = 45 to 65 m s^{-1}.

Winds in Fig. 16.33 are relative to those in the eye. However, the whole tropical cyclone including the eye is often moving. Tropical cyclone **translation speeds** (movement of the center of the storm) can be as slow as M_t = 0 to 5 m s^{-1} as they drift westward in the tropics, and as fast as 25 m s^{-1} as they later move poleward. Average translation speeds of tropical cyclones over the ocean are M_t = 10 to 15 m s^{-1}.

The total wind speed relative to the surface is the vector sum of the translation speed and the rotation speed. On the right quadrant of the storm relative to its direction of movement in the Northern Hemisphere, the translation speed adds to the rotation speed. Thus, tropical cyclone winds are fastest in the tropical cyclone's right quadrant (Fig. 16.34). On the left the translation speed subtracts from the tangential speed, so the fastest total speed in the left quadrant is not as strong as in the right quadrant (Fig. 16.35).

Total speed relative to the surface determines ocean wave and surge generation. Thus, the right quadrant of the storm near the eye wall is most dangerous. Also, tornadoes are likely there.

16.6.3. Radial Velocity

For an idealized tropical cyclone, boundary-layer air is trapped below the top of the boundary layer as winds converge horizontally toward the eye wall. Horizontal continuity in cylindrical coordinates requires:

$$M_{rad} \cdot R = \text{constant} \qquad (16.14)$$

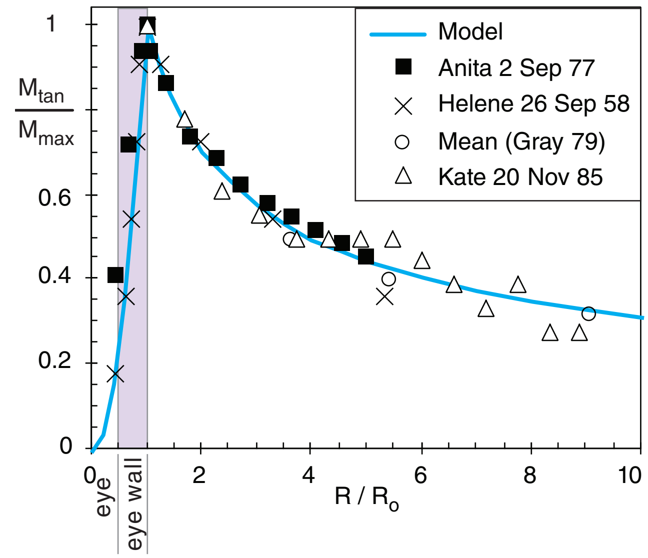


Figure 16.33
Tangential winds near the surface at various radii around tropical cyclones. Model is eq. (16.13).

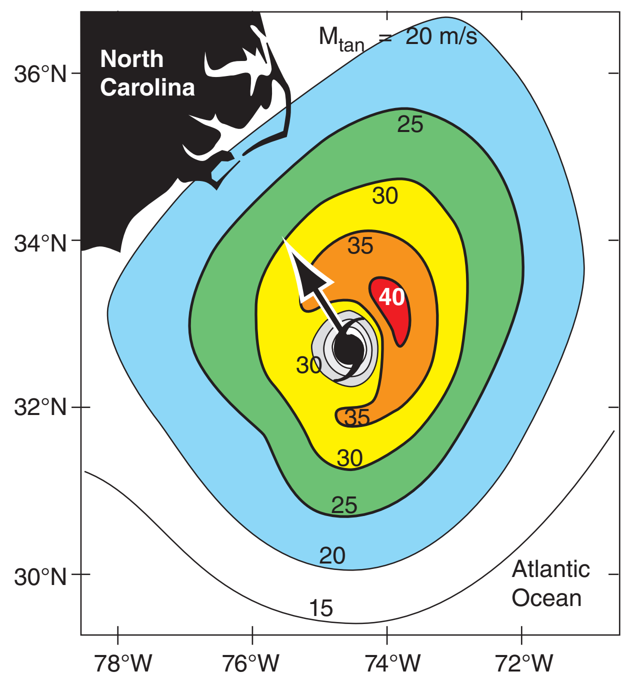


Figure 16.34
Max 1-minute <u>*total*</u> *sustained surface winds (m s^{-1}) around Hurricane Isabel, 0730 UTC 18 Sep 2003. $M_{max\ total}$ = 42 m s^{-1}, and P_{eye} ≈ 95.7 kPa. Arrow shows translation toward 325° at 5.5 m s^{-1}. The Saffir-Simpson Hurricane Wind Scale is based on the fastest total sustained wind (rotational + translational) found anywhere in the hurricane.*

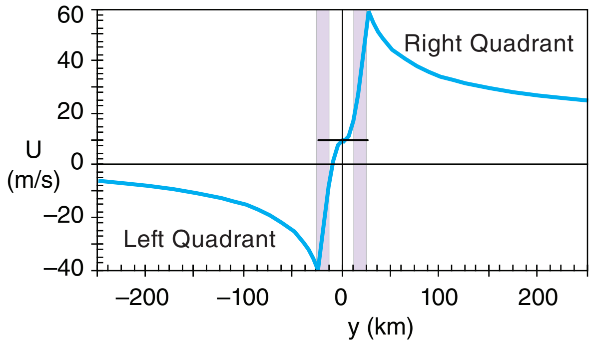


Figure 16.35
Sum of the modeled tangential winds relative to the eye and the translation speed (10 m s^{-1}) of the eye, for a hypothetical tropical cyclone. Assumes M_{max} = 50 m s^{-1} relative to the eye, R_o = 25 km, and uses a coordinate system with the x-axis aligned in the same direction as the translation vector. Purple shading indicates the eye wall locations.

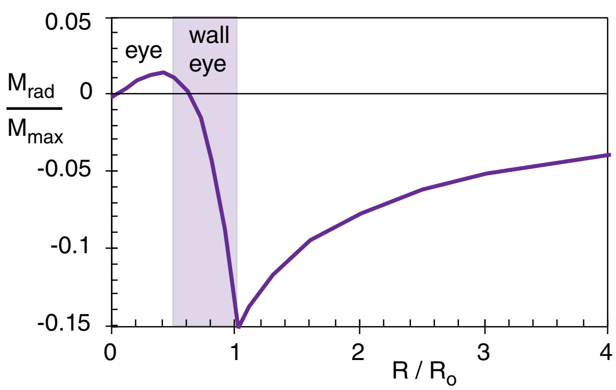


Figure 16.36
Radial winds in a tropical cyclone boundary layer. (Negative values indicate inward motion, converging toward R/R_o = 0.)

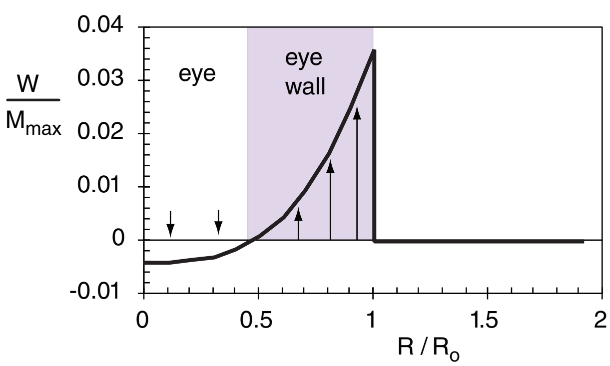


Figure 16.37
Vertical velocity at various radii around tropical cyclones, out of the top of the boundary layer. W_s = –0.2 m s^{-1} in the eye.

where M_{rad} is the radial velocity component, which is negative for inflow. Thus, starting from far outside the tropical cyclone, as R decreases toward R_o, the magnitude of inflow must increase. Inside of R_o, thunderstorm convection removes air mass vertically, implying that horizontal continuity is no longer satisfied.

As distance from the eyewall decreases, surface winds increase, which cause increasing wave height and ocean-surface roughness. The resulting turbulent drag against the ocean surface tends to couple the radial and tangential velocities, which we can approximate by $M_{rad} \propto M_{tan}^2$. Drag-induced inflow such as this eventually converges and forces ascent via the **boundary-layer pumping** process (see the Atmospheric Forces & Winds chapter).

The following equations utilize the concepts above, and are consistent with the tangential velocity in the previous subsection:

(16.15)

$$\frac{M_{rad}}{M_{\max}} = \begin{cases} -\dfrac{R}{R_o}\cdot\left[\dfrac{1}{5}\left(\dfrac{R}{R_o}\right)^3 + \dfrac{1}{2}\dfrac{W_s}{M_{\max}}\dfrac{R_o}{z_i}\right] & \text{for } R \le R_o \\[2ex] -\dfrac{R_o}{R}\cdot\left[\dfrac{1}{5}+\dfrac{1}{2}\cdot\dfrac{W_s}{M_{\max}}\cdot\dfrac{R_o}{z_i}\right] & \text{for } R > R_o \end{cases}$$

where W_s is negative, and represents the average subsidence velocity in the eye. Namely, the horizontal area of the eye, times W_s, gives the total kinematic mass flow downward in the eye. The boundary-layer depth is z_i, and M_{max} is still the maximum tangential velocity.

For example, Fig. 16.36 shows a plot of the equations above, using z_i = 1 km, R_o = 25 km, M_{max} = 50 m s^{-1}, and W_s = –0.2 m s^{-1}. In the eye, subsidence causes air to weakly diverge (positive M_{rad}) toward the eye wall. Inside the eye wall, the radial velocity rapidly changes to inflow (negative M_{rad}), reaching an extreme value of –7.5 m s^{-1} for this case. Outside of the eye wall, the radial velocity smoothly decreases as required by horizontal mass continuity.

16.6.4. Vertical Velocity

At radii less than R_o, the converging air rapidly piles up, and rises out of the boundary layer as thunderstorm convection within the eye wall. The vertical velocity out of the top of the boundary layer, as found from mass continuity, is

(16.16)

$$\frac{W}{M_{\max}} = \begin{cases} \left[\dfrac{z_i}{R_o}\left(\dfrac{R}{R_o}\right)^3 + \dfrac{W_s}{M_{\max}}\right] & \text{for } R < R_o \\[2ex] 0 & \text{for } R > R_o \end{cases}$$

For simplicity, we are neglecting the upward motion that occurs in the spiral rain bands at $R > R_o$.

As before, W_s is negative for subsidence. Although subsidence acts only inside the eye for real tropical cyclones, the relationship above applies it every where inside of R_o for simplicity. Within the eye wall, the upward motion overpowers the subsidence, so our simplification is of little consequence.

Using the same values as for the previous figure, the vertical velocity is plotted in Fig. 16.37. The maximum upward velocity is 1.8 m s^{-1} in this case, which represents an average around the eye wall. Updrafts in individual thunderstorms can be much faster.

Subsidence in the eye is driven by the non-hydrostatic part of the pressure perturbation (Fig. 16.29). Namely, the pressure gradient (shown by the dashed green line between X's in that Fig.) that pushes air upward is weaker than gravity pulling down. This net imbalance forces air downward in the eye.

16.6.5. Temperature

Suppose that the pressure difference between the eye and surroundings at the top of the tropical cyclone is equal and opposite to that at the bottom. From eq. (16.5) the temperature T averaged over the tropical cyclone depth at any radius R is found from:

$$\Delta T(R) = c \cdot \left[\Delta P_{max} - \Delta P(R) \right] \tag{16.17}$$

where c = 1.64 K kPa^{-1}, the pressure difference at the bottom is $\Delta P = P(R) - P_{eye}$, and the temperature difference averaged over the whole tropical cyclone depth is $\Delta T(R) = T_{eye} - T(R)$. When used with eq. (16.11), the result is:

$$\frac{\Delta T}{\Delta T_{max}} = \begin{cases} 1 - \frac{1}{5} \cdot \left(\frac{R}{R_o} \right)^4 & \text{for } R \le R_o \\ \frac{4}{5} \cdot \frac{R_o}{R} & \text{for } R > R_o \end{cases} \tag{16.18}$$

where $\Delta T_{max} = T_{eye} - T_{\infty} = c \cdot \Delta P_{max}$, and c = 1.64 K kPa^{-1}. This is plotted in Fig. 16.38.

16.6.6. Composite Picture

A coherent picture of tropical cyclone structure can be presented by combining all of the idealized models described above. The result is sketched in Fig. 16.39. For real tropical cyclones, sharp cusps in the velocity distribution would not occur because of vigorous turbulent mixing in the regions of strong shear.

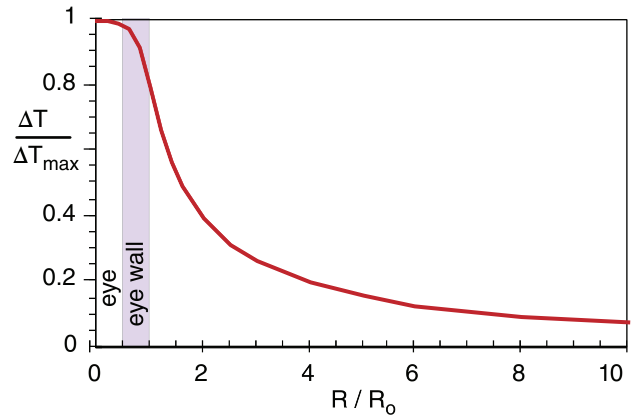


Figure 16.38
Temperature distribution, averaged over a 15 km-thick tropical cyclone, showing the warm core.

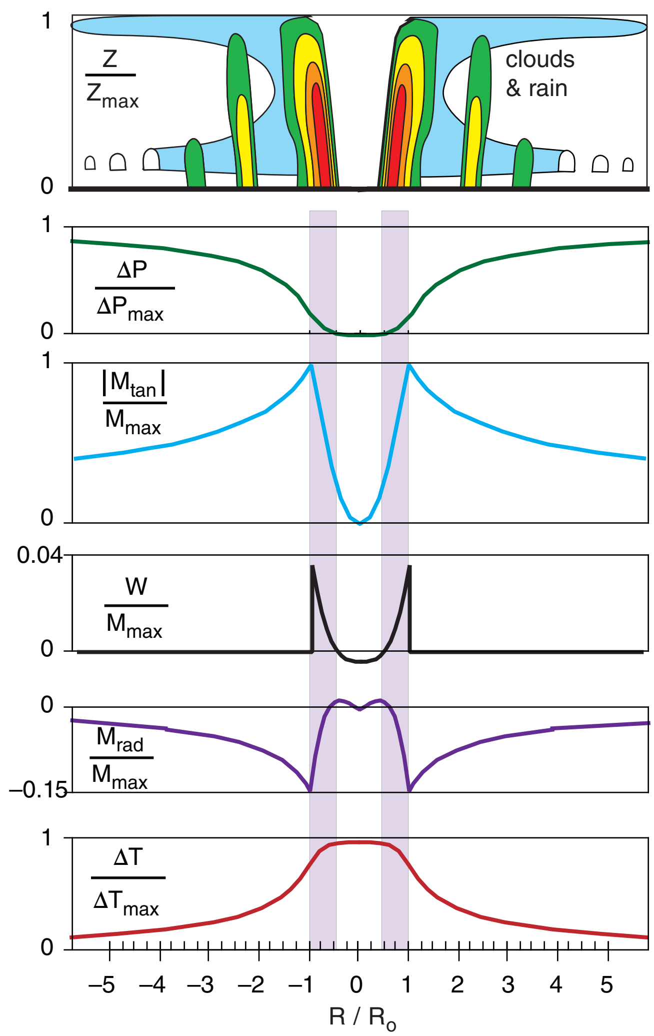


Figure 16.39
Composite tropical cyclone structure based on the idealized model. Pressure differences are at sea level. All horizontal velocities are in the boundary layer, while vertical velocity is across the top of the boundary layer. Temperature differences are averaged over the whole tropical cyclone depth. Purple vertical bands indicate eyewall locations.

Sample Application

A tropical cyclone of critical radius $R_o = 25$ km has a central pressure of 90 kPa. Find the wind components, vertically-averaged temperature excess, and pressure at radius 40 km from the center. Assume $W_s = -0.2$ m s^{-1} in the eye, and $z_i = 1$ km.

Find the Answer

Given: $P_{B\ eye} = 90$ kPa, $R_o = 25$ km, $R = 40$ km, $W_s = -0.2$ m s^{-1}, and $z_i = 1$ km.

Find: $P = ?$ kPa, $T = ?$ °C, $M_{tan} = ?$ m s^{-1}, $M_{rad} = ?$ m s^{-1}, $W = ?$ m s^{-1}

Assume $P_{B\ \infty} = 101.3$ kPa

Figure: Similar to Fig. 16.39. Note that $R > R_o$.

First, find maximum values:

ΔP_{max} = (101.3 – 90kPa) = 11.3 kPa

Use eq. (16.12):

M_{max} = [(20m s^{-1})·kPa$^{-1/2}$]·(11.3kPa)$^{1/2}$ = 67 m s^{-1}

$\Delta T_{max} = c\cdot\Delta P_{max}$=(1.64 K kPa^{-1})·(11.3kPa) = 18.5°C

Use eq. (16.11):

$$\Delta P = (11.3\text{kPa})\cdot\left[1-\frac{4}{5}\cdot\frac{25\text{km}}{40\text{km}}\right] = 5.65\text{ kPa}$$

$P = P_{eye} + \Delta P$ = 90 kPa + 5.65 kPa = **95.65 kPa**.

Use eq. (16.18):

$$\Delta T = (18.5°\text{C})\cdot\frac{4}{5}\cdot\frac{(25\text{km})}{(40\text{km})} = \mathbf{9.25°C}$$

averaged over the whole tropical cyclone depth.

(continues in next column)

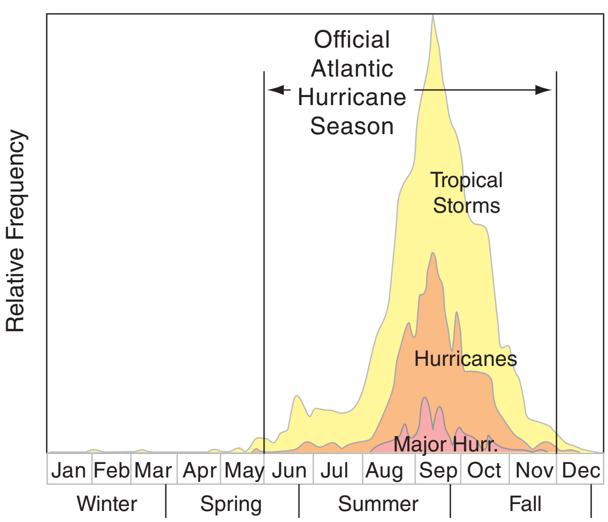


Figure 16.40
Relative frequency vs. time of Atlantic Tropical Storms, Hurricanes, and **Major Hurricanes** *(Saffir-Simpson categories 3 - 5).*

Sample Application *(continuation)*

Use eq. (16.13):

$$M_{tan} = (67\text{m/s})\cdot\sqrt{\frac{25\text{km}}{40\text{km}}} = \mathbf{53\ m\ s^{-1}}$$

Use eq. (16.15):

$$M_{rad} = -(67\text{m/s})\cdot\frac{(25\text{km})}{(40\text{km})}\cdot\left[\frac{1}{5}+\frac{1}{2}\cdot\frac{(-0.2\text{m/s})}{(67\text{m/s})}\cdot\frac{(25\text{km})}{(1\text{km})}\right]$$

$$M_{rad} = \mathbf{-6.8\ m\ s^{-1}}$$

Use eq. (16.16): W = **0 m s^{-1}**

Check: Units OK. Physics OK.

Exposition: Using P_{eye} in Table 16-7, the approximate Saffir-Simpson category of this tropical cyclone is borderline between levels 4 and 5, and thus is very intense.

16.7. CLIMATOLOGY

16.7.1. Seasonality

Tropical cyclones are most frequent in late Summer and Fall of their respective hemisphere. This is because the sun has been at its highest in the sky, causing the top layers of the tropical ocean to accumulate the most heat. Fig. 16.40 shows the frequency of Atlantic Hurricanes vs. time. Table 16-6 shows periods of frequent tropical cyclones in all the ocean basins.

16.7.2. Locations of Strongest Cyclones

The largest number of strongest tropical cyclones is in the northwestern Pacific. The reason is that the Pacific is a larger ocean with warmer sea-surface

Table 16-6. Tropical Cyclone Seasons. Start & end dates are for the most active period of the storm season, but some storms occur outside of this peak season.

Location	Start	Peak	End
Atlantic	1 June	mid Sep	30 Nov
NE Pacific	15 May	late Aug /early Sep	30 Nov
NW Pacific*	1 July	late Aug /early Sep	30 Nov
N. Indian	1 Apr	2 peaks: Apr-Jun, & late Sep-early Dec	31 Dec
S. Indian	1 Oct	2 peaks: mid-Jan, & mid Feb - early Mar	31 May
Australia & SW Pacific	1 Oct	late Feb - early Mar	31 May

** The NW Pacific has typhoons all year.*

temperatures, allowing typhoons more opportunity to organize and strengthen. Activity ranges from 17 typhoons in a slow year (1998) to 35 in an active year (1971). Also, the larger **fetch** (distance that wind blows over the ocean) allows larger ocean waves, which can cause more destruction (and better surfing further away).

The opposite extreme is the South Atlantic, which has had only 2 recorded tropical cyclones in the past century. One was Cyclone Catarina, which struck Brazil in March 2004. The other was a Tropical Storm that formed west of Congo in April 1991. There may have been other tropical cyclones in the South Atlantic that were not recorded historically.

There are two reasons for the dearth of tropical cyclones in the South Atlantic. One is the weaker and sometimes nonexistent ITCZ, which reduces tropical cyclone triggering because of less convergence and less initial vorticity. The other is strong wind shear in the upper troposphere, which rips apart thunderstorm clusters before they can become tropical cyclones.

16.7.3. Natural Cycles & Changes in Activity

Atlantic hurricanes have a very large natural variability from year to year. For example, in the Atlantic there were only 4 hurricanes recorded in 1983, and 19 in 1994. An active hurricane year was 2005, with 14 hurricanes and 13 other tropical storms.

Hidden behind this large annual variability are longer-time-period variations of weaker amplitude, making them more difficult to detect and confirm. One is a **natural 40-year cycle** in Atlantic hurricane power (based on wind-speed cubed accumulated over the lifetime of all Atlantic hurricanes). This power was relatively high during the 1950s and 1960s, and was weaker during the 1970s, 1980s, and early 1990s. Since the late 1990s and 2000s, hurricane power has increased again.

El Niño/La Niña, an irregular 3 to 5 year cycle (see the Natural Climate Processes chapter), also causes long time-scale variations in hurricane activity. During El Niño, there is a tendency for reduced tropical cyclone activity, and some displacement of these storms closer to the equator. The reason is that stronger west winds aloft during El Niño cause stronger wind shear across the troposphere, thereby inhibiting tropical cyclogenesis. Conversely, tropical cyclone activity is enhanced during La Niña.

Of concern these days is human-caused **global warming**. While most scientists suspect that there will be some effect on tropical cyclones if global warming continues, they have not yet found a clear signal. Debate continues.

16.8. HAZARDS

16.8.1. Human Population and Zoning

The most important factor causing increased deaths and destruction from tropical cyclones is the increase in **global population**. With population growth, more people live in coastal areas, which are seen as desirable in spite of the threat of tropical cyclones. Even if tropical-cyclone activity were to remain relatively constant, the impacts on humans would increase as population increases.

As more structures are built in vulnerable areas, so increase the property losses caused by tropical cyclone destruction. It also becomes more difficult to evacuate people along inadequate transportation networks. As more people move from farms to cities, there are increased fatalities due to urban flooding caused by heavy tropical-cyclone rainfall.

In highly developed countries, an easy solution would be proper land-use **zoning**. Namely, governments would not let people live and work in threatened areas. Instead, these areas would be used for parks, floodable farmland, wildlife refuges, etc. However, zoning is a political activity that sometimes results in poor decisions in the face of pressure from real-estate developers who want to build more waterfront properties. An example of a questionable decision is the reconstruction of New Orleans, Louisiana, USA, at its original location after being destroyed by hurricane Katrina in 2005.

Also, people are unfortunately encouraged to live in threatened areas because of the existence of hurricane **insurance** and hurricane **disaster relief**. Namely, some individuals choose to make these poor decisions on where to live because they do not have to bear the full costs of reconstruction — instead the cost is borne by all taxpayers.

In less developed, highly populated, low-lying countries such as Bangladesh, an additional problem is an inadequate **warning** system. Even when tropical cyclone tracks are successfully predicted, sometimes the warning does not reach rural people, and often there is inadequate transportation to enable their evacuation. Bangladesh has suffered terribly from tropical cyclones: 500,000 deaths in Nov 1970, over 11,000 deaths in May 1985 , and 150,000 deaths in Apr 1991.

These aspects of tropical-storm hazards are therefore social (cultural, political, religious, etc.). It cannot be assumed that all problems can be ameliorated by technical solutions (more dams; higher levies). Instead, some tough decisions need to be made on zoning, transportation, and population growth.

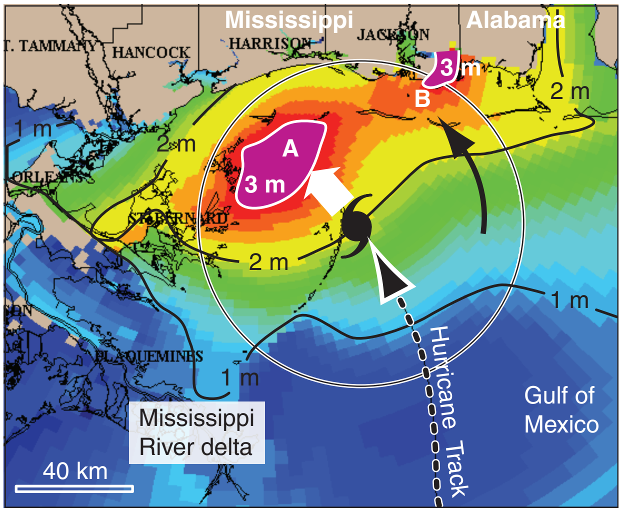


Figure 16.41
Simulated storm surge associated with a hypothetical landfalling hurricane. Color shading and contours indicate rise (m) in sea level. (Modified from a NOAA/HRD figure).

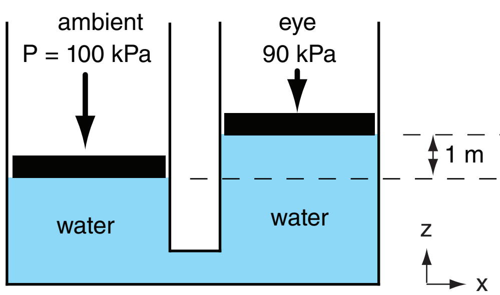


Figure 16.42
Reduced atmospheric pressure in the eye allows sea-level to rise.

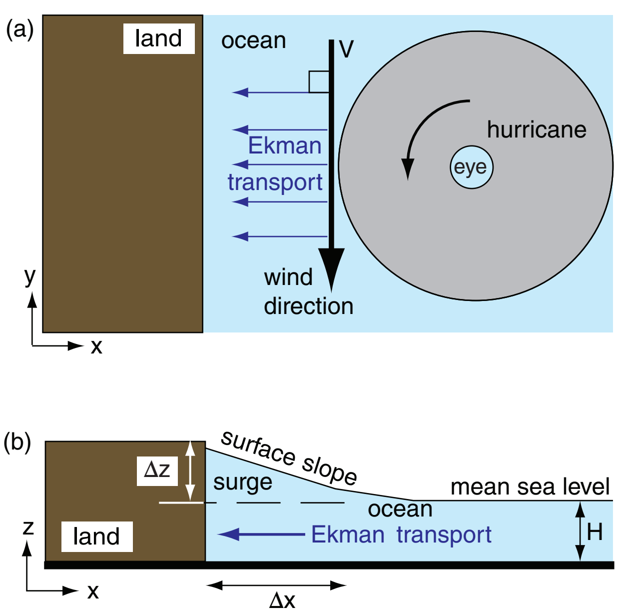


Figure 16.43
Onshore Ekman transport prior to tropical cyclone landfall, in the Northern Hemisphere. (a) Top view. (b) View from south.

16.8.2. Storm Surge

Much of the tropical cyclone-caused damage results from inundation of coastal areas by high seas (Fig. 16.41). The rise in sea level (i.e., the **storm surge**) is caused by the reduced atmospheric pressure in the eye, and by wind blowing the water against the coast to form a large propagating surge called a Kelvin wave. Table 16-7 gives typical storm-surge heights. High tides and high surface-waves can exacerbate the damage.

16.8.2.1. Atmospheric Pressure Head

In the eye of the tropical cyclone, atmospheric surface pressure is lower than ambient. Hence the force per unit area pushing on the top of the water is less. This allows the water to rise in the eye until the additional head (weight of fluid above) of water compensates for the reduced head of air (Fig. 16.42).

The amount of rise Δz of water in the eye is

$$\Delta z = \frac{\Delta P_{max}}{\rho_{liq} \cdot |g|} \qquad \bullet(16.19)$$

where $|g|$ is gravitational acceleration magnitude (9.8 m s^{-2}), ρ_{liq} = 1025 kg m^{-3} is the density of sea water, and ΔP_{max} is the atmospheric surface pressure difference between the eye and the undisturbed environment.

To good approximation, this is

$$\Delta z \approx a \cdot \Delta P_{max} \qquad (16.20)$$

where a = 0.1 m kPa^{-1}. Thus, in a strong tropical cyclone with eye pressure of 90 kPa (causing $\Delta P_{max} \approx$ 10 kPa), the sea level would rise 1 m.

16.8.2.2. Ekman Transport

Recall from the General Circulation chapter that ocean currents are generated by wind drag on the sea surface. The Ekman spiral describes how the current direction and speed varies with depth. The net **Ekman transport**, accumulated over all depths, is exactly perpendicular to the surface wind direction.

In the Northern Hemisphere, this net transport of water is to the right of the wind, and has magnitude:

$$\frac{Vol}{\Delta t \cdot \Delta y} = \frac{\rho_{air}}{\rho_{water}} \cdot \frac{C_D \cdot V^2}{f_c} \qquad (16.21)$$

where Vol is the volume of water transported during time interval Δt, Δy is a unit of length of coastline parallel to the mean wind, V is the wind speed near the surface (actually at 10 m above the surface), C_D is the drag coefficient, f_c is the Coriolis parameter, ρ_{air}

= 1.225 $kg{\cdot}m^{-3}$ is the air density at sea level, and ρ_{water} = 1025 $kg{\cdot}m^{-3}$ is sea-water density.

As a tropical cyclone approaches the eastern coast of continents, the winds along the front edge of the tropical cyclone are parallel to the coast, from north to south in the Northern Hemisphere. Hence, there is net Ekman transport of water directly toward the shore, where it begins to pile up, creating a storm surge (Fig. 16.43).

If the tropical cyclone were to hover just offshore for sufficient time to allow a steady-state condition to develop, then the Ekman transport toward the shore would be balanced by downslope sloshing of the surge. The surge slope $\Delta z/\Delta x$ for that hypothetical equilibrium is:

$$\frac{\Delta z}{\Delta x} \approx \frac{\rho_{air}}{\rho_{water}} \cdot \frac{C_D \cdot V^2}{|g| \cdot H} \qquad (16.22)$$

where H is the unperturbed ocean depth (e.g., 50 m) near the coast, and the other variables are the same as for eq. (16.21).

16.8.2.3. Kelvin Wave

Because the tropical cyclone has finite size, Ekman transport is localized to the region immediately in front of the tropical cyclone. Thus, water is piled higher between the tropical cyclone and the coast than it is further north or south along the coast. When viewed from the East, the surge appears as a long-wavelength wave, called a **Kelvin wave** (Fig. 16.44).

The propagation speed of the wave, called the **phase speed** c, is

$$c = \sqrt{|g| \cdot H} \qquad \bullet(16.23)$$

which is also known as the **shallow-water wave speed**, where $|g|$ is gravitational acceleration magnitude and H is average water depth. Typical phase speeds are 15 to 30 m s^{-1}. These waves always travel with the coast to their right in the Northern Hemisphere, so they propagate southward along the east coast of continents, and northward along west coasts.

As the wave propagates south along the East Coast, it will hug the coast and inundate the shore immediately south of the tropical cyclone. Meanwhile, Ekman transport continues to build the surge in the original location. The net result is a continuous surge of high water along the shore that is closest to, and south of, the tropical cyclone. Typical surge depths can be 2 to 10 m at the coast, with extreme values of 13 to 20 m.

Table 16-7. Storm surge height S and sea-level pressure P_{eye}, from the old Saffir-Simpson classification. CAUTION: Actual storm-surge heights can vary significantly from these typical values. This is one of the reasons why the new Saffir-Simpson scale doesn't use S.

Category	P_{eye} (kPa)	S (m)
1	≥ 98.0	1.2 - 1.6
2	97.9 - 96.5	1.7 - 2.5
3	96.4 - 94.5	2.6 - 3.9
4	94.4 - 92.0	4.0 - 5.5
5	< 92.0	> 5.5

Sample Application

For tropical cyclone-force winds of 40 m s^{-1}, over a continental shelf portion of ocean of depth 50 m, find the volume transport rate and equilibrium surge slope. Assume C_D = 0.01 .

Find the Answer

Given: ρ_{air} = 1.225 $kg{\cdot}m^{-3}$, ρ_{water} = 1025 $kg{\cdot}m^{-3}$
M = 40 m s^{-1}, H = 50 m, C_D = 0.01
Find: $Vol/(\Delta t{\cdot}\Delta y)$ = ? $m^2\ s^{-1}$, and $\Delta z/\Delta x$ = ?
Assume: f_c = 0.00005 s^{-1}

Use eq. (16.21):

$$\frac{Vol}{{}_{s^{-1}}\Delta t \cdot \Delta y} = \frac{(1.225\text{kg/m}^3)}{(1025\text{kg/m}^3)} \cdot \frac{(0.01)\cdot(40\text{m/s})^2}{(0.00005\text{s}^{-1})} = \underline{382\ \text{m}^2}$$

Use eq. (16.22):

$$\frac{\Delta z}{\Delta x} \approx \frac{(1.225\text{kg/m}^3)}{(1025\text{kg/m}^3)} \cdot \frac{(0.01)\cdot(40\text{m/s})^2}{(9.8\text{m/s}^2)\cdot(50\text{m})} = \underline{3.9\text{x}10^{-5}}$$

Check: Units OK. Physics OK.
Exposition: The flow of water is tremendous. As it starts to pile up, the gradient of sea-level begins to drive water away from the surge, so it does not continue growing. The slope corresponds to 4 cm rise per km distance toward the shore. Over 10s to 100s km, the rise along the coast can be significant.

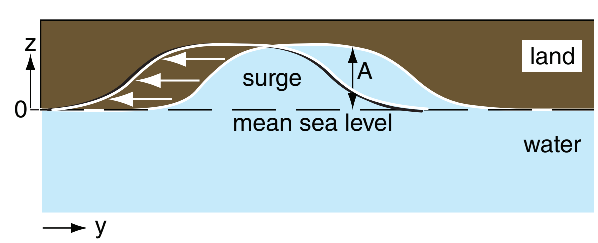


Figure 16.44
The surge, viewed from the east towards vertical cliffs of land, is a Kelvin wave that propagates south. For low-lying coastal areas with no cliffs, the Kelvin-wave storm surge would spread inland and inundate (flood) the coastal areas.

Sample Application

Using info from the previous Sample Application, find the Kelvin wave phase speed, and the growth rate if the tropical cyclone follows the wave southward.

Find the Answer

Given: (same as previous example)
Find: $c = ?\ \text{m s}^{-1}$, $\Delta A/\Delta t = ?\ \text{m s}^{-1}$

Use eq. (16.23):

$$c = \sqrt{(9.8\text{m/s}^2)\cdot(50\text{m})} = \underline{\mathbf{22.1\ m\ s^{-1}}}$$

Use eq. (16.24):

$$\frac{\Delta A}{\Delta t} \approx \frac{(1.225\text{kg/m}^3)}{(1025\text{kg/m}^3)} \cdot \frac{(0.01)\cdot(40\text{m/s})^2}{\sqrt{(9.8\text{m/s}^2)\cdot(50\text{m})}}$$
$$= \underline{\mathbf{0.00086\ m\ s^{-1}}}$$

Check: Units OK. Physics OK.
Exposition: Tropical cyclones usually turn northward. If tropical cyclones were to translate southward with speed 22.1 m s^{-1}, matching the Kelvin wave speed, then an exceptionally dangerous situation would develop with amplification of the surge by over 3 m h^{-1}.

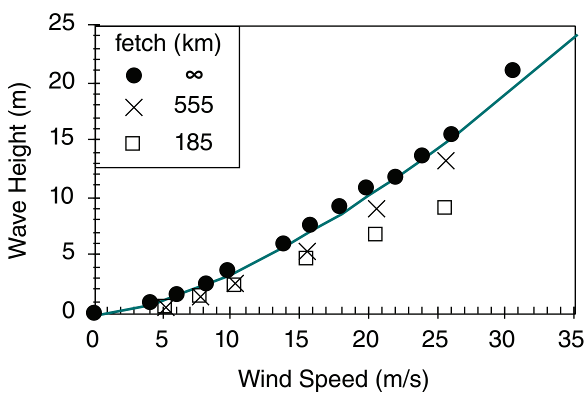


Figure 16.45
Wave height of wind-generated surface waves. Solid line is eq. (16.25) for unlimited fetch and duration. Data points are ocean observations with different fetch.

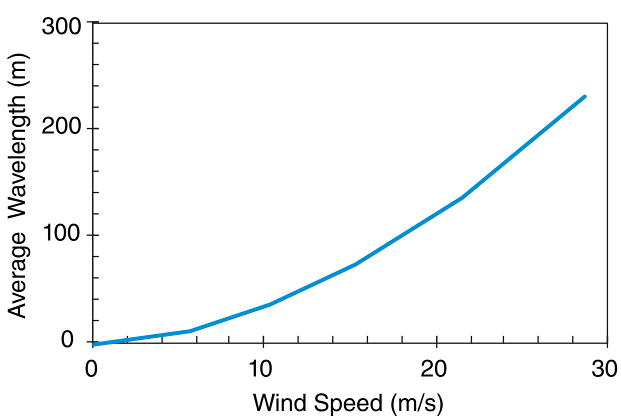


Figure 16.46
Wavelength of wind waves. Solid line is eq. (16.26).

Should the tropical cyclone move southward at a speed nearly equal to the Kelvin wave speed, then Ekman pumping would continue to reinforce and build the surge, causing the amplitude of the wave A to grow according to:

$$\frac{\Delta A}{\Delta t} \approx \frac{\rho_{air}}{\rho_{water}} \cdot \frac{C_D \cdot V^2}{\sqrt{|g| \cdot H}} \qquad (16.24)$$

where the amplitude A is measured as maximum height of the surge above mean sea level, and the other variables are the same as for eq. (16.21).

16.8.3. Surface Wind-waves

Waves are generated on the sea surface by action of the winds. Greater winds acting over longer distances (called **fetch**) for greater time durations can excite higher waves. High waves caused by tropical cyclone-force winds are not only a hazard to shipping, but can batter structures and homes along the coast.

Four coastal hazards of a tropical cyclone are:

- wave scour of the beach under structures,
- wave battering of structures,
- surge flooding, and
- wind damage.

The first two hazards exist only right on the coast, in the beach area. But the inundation (flooding) from the storm surge can reach 10 to 15 km inland from the coast, depending on the slope of the land.

For wind speeds M up to tropical cyclone force, the maximum-possible wave height (for unlimited fetch and duration) can be estimated from:

$$h = h_2 \cdot \left(\frac{M}{M_2}\right)^{3/2} \qquad (16.25)$$

where $h_2 = 4$ m and $M_2 = 10$ m s^{-1}. Wave heights are plotted in Fig. 16.45.

As winds increase beyond tropical cyclone force, the wave tops become partially chopped off by the winds. Thus, wave height does not continue to increase according to eq. (16.25). For extreme winds of 70 m s^{-1}, the sea surface is somewhat flat, but poorly defined because of the mixture of spray, foam, and chaotic seas that appear greenish white during daytime.

Wavelengths of the wind-waves also increase with wind speed. Average wavelengths λ can be approximated by:

$$\lambda = \lambda_2 \left(\frac{M}{M_2}\right)^{1.8} \qquad (16.26)$$

where $\lambda_2 = 35$ m, and $M_2 = 10$ m s^{-1}. Wavelengths are plotted in Fig. 16.46.

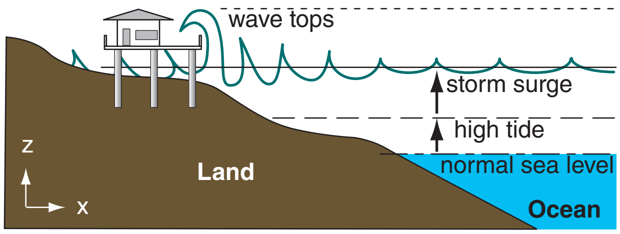


Figure 16.47
Tides, storm surge, and waves are additive.

The longest wavelength waves are called swell, and can propagate large distances, such as across whole oceans. Hence, a tropical cyclone in the middle tropical Atlantic can cause large surf in Florida well before the storm reaches the coast.

Wind and waves as affect mariners are classified according to the Beaufort scale. The INFO Box with Table 16-8 gives a historical description of the Beaufort scale. A modern description is in Table 16-9.

Tides, storm surges, and wind waves are additive (Fig. 16.47). Namely, if the storm surge and high waves happen to occur during high tide according to routine astronomic tide tables, then the coastal destruction is likely to be greatest.

For safety, houses at coasts threatened with storm surges and tsunami are usually built on top of concrete or steel piles. These piles are driven very deep into the land to survive wave scour and erosion of the land. While the floor deck is above expected storm-surge plus high tide levels in this figure, the waves can still batter and damage the structure.

Sample Application

Find the maximum possible wave height (assuming unlimited fetch) and wavelength for tropical cyclone force winds of 35 m s^{-1}.

Find the Answer

Given: M = 35 m s^{-1}
Find: h = ? m, λ = ? m

Use eq. (16.25):

$$h = (4\text{m})\cdot\left(\frac{35\text{m/s}}{10\text{m/s}}\right)^{3/2} = \underline{26.2\text{ m}}$$

Use eq. (16.26):

$$\lambda = (35\text{m})\cdot\left(\frac{35\text{m/s}}{10\text{m/s}}\right)^{1.8} = \underline{334\text{ m}}$$

Check: Units OK. Physics OK. Agrees with Figs.
Exposition: Wavelengths are much longer than wave heights. Thus, wave slopes are small — less that 1/10. Only when these waves reach shore does wave slope grow until the waves break as surf.

INFO • Beaufort Wind-Scale History

In 1805, Admiral Beaufort of the British Navy devised a system to estimate and report wind speeds based on the amount of canvas sail that a full-rigged frigate could carry. It was updated in 1874, as listed in Table 16-8 for historical interest. Modern descriptors for the Beaufort wind scale are in Table 16-9.

Table 16-8. Legend: B = Beaufort number; D = modern classification; M = wind speed in knots (2 knots ≈ 1 m s^{-1}), S1 = speed through smooth water of a well-conditioned man-of-war carrying all sails un-reefed (full), sailing close to the wind; S2 = un-reefed (full) sails that a well-conditioned man-of-war could just carry in chase, sailing close to the wind; S3 = sails that a well-conditioned man-of-war could scarcely bear. [A reefed sail exposes less than its full area to the wind.]

B	D	M (kt)	Deep Sea Criteria
0	Calm	0 - 1	S1 = Becalmed
1	Light Air	1 - 3	S1 = Just sufficient to give steerageway
2	Slight Breeze	4 - 6	S1 = 1 - 2 knots
3	Gentle Breeze	7 - 10	S1 = 3 - 4 knots
4	Moderate Breeze	11 - 16	S1 = 5 - 6 knots
5	Fresh Breeze	17 - 21	S2 = Royals, etc.
6	Strong Breeze	22 - 27	S2 = Topgallant sails
7	High Wind	28 - 33	S2 = Topsails, jib, etc.
8	Gale	34 - 40	S2 = Reefed upper topsails and courses
9	Strong Gale	41 - 48	S2 = Lower topsails and courses
10	Whole Gale	49 - 55	S3 = lower main topsail and reefed foresail
11	Storm	56 - 65	S3 = storm staysails
12	Hurricane	> 65	S3 = no canvas

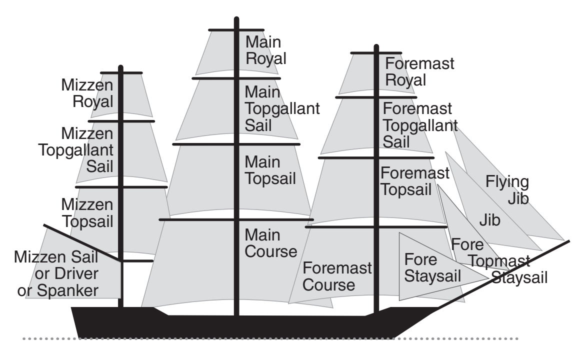


Figure 16.d. *Frigate sails.*

Table 16-9. Beaufort wind scale. B = Beaufort Number. (See INFO Box for historical info.)

B	Descrip-tion	Wind Speed (km/h)	Wind Speed (m/s)	Wave Height (m)	Sea Conditions (in deep ocean)	Land Conditions
0	Calm	< 1	< 0.3	0	Flat.	Calm. Smoke rises vertically.
1	Light Air	1 - 5	0.3 - 1.5	0 -00.2	Ripples without crests.	Wind motion visible in smoke.
2	Light Breeze	6 - 11	1.5 - 3.3	0.2 - 0.5	Small wavelets. Crests of glassy appearance, not breaking.	Wind felt on exposed skin. Leaves rustle.
3	Gentle Breeze	12 - 19	3.3 - 5.5	0.5 - 1	Large wavelets. Crests begin to break; scattered whitecaps.	Leaves and smaller twigs in constant motion.
4	Moderate Breeze	20 - 28	5.5 - 8.0	1 - 2	Small waves with breaking crests. Fairly frequent whitecaps.	Dust and loose paper raised. Small branches begin to move.
5	Fresh Breeze	29 - 38	8.0 - 11	2 - 3	Moderate waves of some length. Many whitecaps. Small amounts of spray.	Branches of a moderate size move. Small trees begin to sway.
6	Strong Breeze	39 - 49	11 - 14	3 - 4	Long waves begin to form. White foam crests are very frequent. Some airborne spray is present.	Large branches in motion. Whistling heard in overhead wires. Umbrella use becomes difficult. Empty plastic garbage cans tip over.
7	High Wind, Moderate Gale, Near Gale	50 - 61	14 - 17	4 - 5.5	Sea heaps up. Some foam from breaking waves is blown into streaks along wind direction. Moderate amounts of airborne spray.	Whole trees in motion. Effort needed to walk against the wind. Swaying of skyscrapers may be felt, especially by people on upper floors.
8	Gale, Fresh Gale	62 - 74	17 - 20	5.5 - 7.5	Moderately high waves with breaking crests forming spindrift. Well-marked streaks of foam are blown along wind direction. Considerable airborne spray.	Some twigs broken from trees. Cars veer on road. Progress on foot is seriously impeded.
9	Strong Gale	75 - 88	21 - 24	7 - 10	High waves whose crests sometimes roll over. Dense foam is blown along wind direction. Large amounts of airborne spray may begin to reduce visibility.	Some branches break off trees, and some small trees blow over. Construction/temporary signs and barricades blow over. Damage to circus tents and canopies.
10	Storm, Whole Gale	89 - 102	25 - 28	9 - 12.5	Very high waves with overhanging crests. Large patches of foam from wave crests give the sea a white appearance. Considerable tumbling of waves with heavy impact. Large amounts of airborne spray reduce visibility.	Trees are broken off or uprooted, saplings bent and deformed. Poorly attached asphalt shingles and shingles in poor condition peel off roofs.
11	Violent Storm	103 - 117	29 - 32	11.5 - 16	Exceptionally high waves. Very large patches of foam, driven before the wind, cover much of the sea surface. Very large amounts of airborne spray severely reduce visibility.	Widespread vegetation damage. Many roofing surfaces are damaged; asphalt tiles that have curled up and/or fractured due to age may break away completely.
12	Hurricane	≥ 118	≥ 33	≥14	Huge waves. Sea is completely white with foam and spray. Air is filled with driving spray, greatly reducing visibility.	Very widespread damage to vegetation. Some windows may break; mobile homes and poorly constructed sheds and barns are damaged. Debris may be hurled about.

16.8.4. Inland Flooding

In developed countries such as the USA, storm-surge warning and evacuation systems are increasingly successful in saving lives of coastal dwellers. However, inland flooding due to heavy rain from decaying tropical storms is increasingly fatal — causing roughly 60% of the tropical cyclone-related deaths in the USA during the past 30 years.

Streams and storm drains overflow, trapping people in cars and on roof tops. For this reason, people should not be complacent about former tropical cyclones reaching their inland homes, because these dying tropical cyclones contain so much tropical moisture that they can cause record-setting rainfalls.

The inland flooding hazard can affect people hundreds of kilometers from the coast. Of the 56 people who died in Hurricane Floyd (1999) in the eastern USA, 50 drowned in inland floods caused by heavy rains. Tropical storm Alberto dropped 53 cm of rain over Americus, Georgia, where 33 people drowned in 1994. Over 200 people drowned in Pennsylvania, New York, and New England from Hurricane Diane in 1955.

More recently, many unnecessary drowning fatalities have been caused by people driving into water of unknown depth covering the road. For people comfortable in driving their usual roads day after day, many find it hard to believe that these roads can become impassable. They unknowingly drive into deep water, causing the engine to stall and the car to stop in the middle of the water. If the water continues to rise, the car and passengers can be carried away. An easy solution is to approach each flooded road with caution, and be prepared to interrupt your journey and wait until the flood waters subside.

16.8.5. Thunderstorms, Lightning & Tornado Outbreaks

Because tropical cyclones are made of thunderstorms, they contain all the same hazards as thunderstorms. These include lightning, downbursts, gust fronts, downpours of rain, and tornadoes (see the Thunderstorm chapter for hazard details). Lightning is somewhat infrequent in the eyewall (about 12 cloud-to-ground strikes per hour), compared to about 1000 strikes per hour for midlatitude MCSs.

Tropical cyclones can cause **tornado outbreaks**. For example, Hurricanes Cindy and Katrina each spawned 44 tornadoes in the USA in 2005. Hurricane Rita spawned 101 tornadoes in Sep 2005. Hurricanes Frances and Ivan spawned 103 and 127 tornadoes, respectively, in Sep 2004.

Most tornadoes are weak (≤ EF2), and occur in the right front quadrant of tropical cyclones in the N. Hemisphere. Although some form near the eyewall, most tornadoes form in the thunderstorms of outer rain bands 80 to 480 km from the cyclone center.

16.9. TROPICAL CYCLONE FORECASTING

16.9.1. Prediction

The most important advance in tropical cyclone prediction is the weather satellite (see the Satellites & Radar chapter). Satellite images can be studied to find and track tropical disturbances, depressions, storms, and cyclones. By examining loops of sequential images of tropical-cyclone position, their past track and present translation speed and direction can be determined. Satellites can be used to estimate rainfall intensity and storm-top altitudes.

Research aircraft (**hurricane hunters**) are usually sent into dangerous storms to measure pressure, wind speed, temperature, and other variables that are not easily detected by satellite. Also, they can fly transects through the middle of the tropical storm to precisely locate the eye (Fig. 16.48). The two organizations that do this for Atlantic Hurricanes are the US Air Force Reserves **53rd Reconnaissance Squadron**, and the Aircraft Operations Center of the US **National Oceanic and Atmospheric Administration** (**NOAA**).

Forecasting future tracks and intensity of tropical cyclones is more difficult, and is prone to error. Computer codes (called models) describing atmospheric physics and dynamics [see the Numerical Weather Prediction (NWP) chapter] are run to fore-

Figure 16.48
Photo inside the eye of Hurricane Rita on 21 Sep 2005, taken from a NOAA P3 aircraft. Rita was category 5 at max intensity. (Image courtesy of NOAA/AOC.)

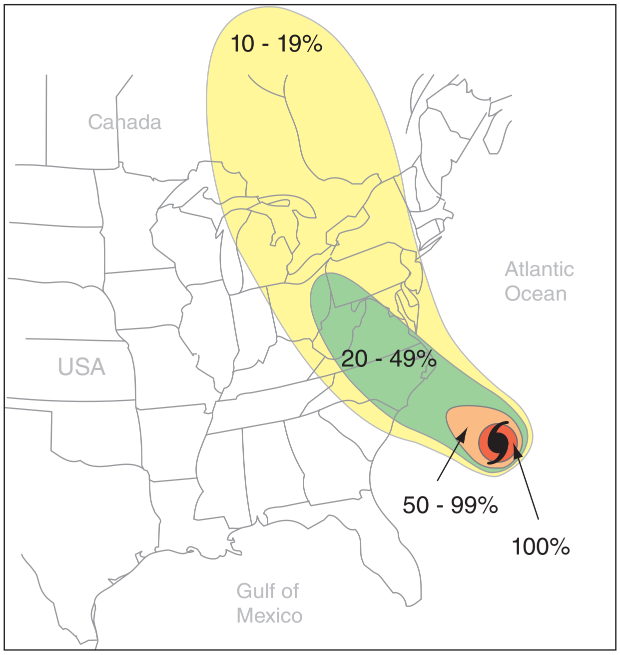


Figure 16.49
Probability forecast that the eye of Hurricane Isabel would pass within 120 km of any point on the map during the 72 h starting 1800 UTC on 17 Sep 2003. The actual location of the hurricane at this time is indicated with the hurricane map symbol.

cast the weather. But different models yield different forecasts of hurricane tracks. Human forecasters therefore consider all available NWP model forecasts, and issue probability forecasts (Fig. 16.49) on the likelihood that any tropical cyclone will strike different sections of coastline. Local government officials and emergency managers then make the difficult (and costly) decision on whether to evacuate any sections of coastline.

Predicting tropical cyclone intensity is even more difficult. Advances have been made based on sea-surface temperature measurements, atmospheric static stability, ambient wind shear, etc. Also, processes such as eyewall replacement and interaction with other tropical and extratropical systems are considered. But much more work needs to be done.

Different countries have their own organizations to issue forecasts. In the USA, the responsible organization is the **National Hurricane Center**, also known as the **Tropical Prediction Center**, a branch of NOAA. This center issues the following forecasts:

- **Tropical Storm Watch** - tropical storm conditions are possible within 36 h for specific coastal areas. (Includes tropical storms as well as the outer areas of tropical cyclones.)
- **Tropical Storm Warning** - tropical storm conditions are expected within 12 h or less for specific coastal areas. (Includes tropical storms as well as the outer areas of tropical cyclones.)
- **Hurricane Watch** - hurricane conditions are possible within 36 h for specific coastal areas.
- **Hurricane Warning** - hurricane conditions are expected within 24 h or less for specific coastal areas. Based on the worst of expected winds, high water, or waves.

In Canada, the responsible organization is the **Canadian Hurricane Center,** a branch of the **Meteorological Service of Canada**.

16.9.2. Safety

The US National Hurricane Center offers these recommendations for a Family Disaster Plan:

- Discuss the type of hazards that could affect your family. Know your home's vulnerability to storm surge, flooding and wind.
- Locate a safe room or the safest areas in your home for each hurricane hazard. In certain circumstances the safest areas may not be in your home but could be elsewhere in your community.
- Determine escape routes from your home and places to meet. These should be measured in tens of miles rather than hundreds of miles.
- Have an out-of-state friend as a family contact, so all your family members have a single point of contact.
- Make a plan now for what to do with your pets if you need to evacuate.
- Post emergency telephone numbers by your phones and make sure your children know how and when to call 911.
- Check your insurance coverage — flood damage is not usually covered by homeowners insurance.
- Stock non-perishable emergency supplies and a Disaster Supply Kit.
- Use a NOAA weather radio. Remember to replace its battery every 6 months, as you do with your smoke detectors.
- Take First Aid, cardiopulmonary resuscitation (**CPR**) and disaster-preparedness classes.

If you live in a location that receives an evacuation order, it is important that you follow the instructions issued by the local authorities. Some highways are designated as evacuation routes, and traffic might be rerouted to utilize these favored routes. It is best to get an early start, because the roads are increasingly congested due to population growth. Before you evacuate, be sure to board up your home to protect it from weather and looters.

Many people who live near the coast have storm shutters permanently installed on their homes, which they can close prior to tropical cyclone arrival. For windows and storefronts too large for shutters, have large pieces of plywood on hand to screw over the windows.

If a tropical cyclone overtakes you and you cannot escape, stay indoors away from windows. Street signs, corrugated metal roofs, and other fast-moving objects torn loose by tropical cyclone-force winds can slice through your body like a guillotine.

16.10. REVIEW

Hurricanes are tropical cyclones. They have low-pressure centers, called eyes, and rotation is cyclonic (counterclockwise in the Northern Hemisphere) near the surface. The tropical cyclone core is warm, which causes high pressure to form in the eye near the top of the storm. This high pressure drives diverging, anticyclonic winds out of the tropical cyclone.

Tropical cyclones are born over tropical oceans with temperature ≥26.5°C over 50 m or more depth. Evaporation from the warm ocean into the windy boundary layer increases the energy in the storm, which ultimately drives a circulation similar to a Carnot-cycle heat engine. Tropical cyclones die over cold water and over land, not due to the extra drag caused by buildings and trees, but due mostly to the lack of strong evaporation from the ocean.

Because tropical cyclones are born in the trade-wind regions of the global circulation, they are initially blown westward. Many eventually reach the eastern shores of continents where the global circulation turns them poleward. At the equator there is no Coriolis force; hence there is no rotation available to be concentrated into tropical cyclones. Thus, tropical cyclones are most likely to form between 10° to 30° latitude during autumn.

Updrafts are strongest in the eye wall of thunderstorms encircling the clear eye. Rotation is initially gathered from the absolute angular momentum associated with the Earth's rotation. As the storm develops and gains speed, centrifugal force dominates over Coriolis force within about 100 km of the eye, causing winds that are nearly cyclostrophic in the bottom third of the troposphere. Simple analytical models can be built to mimic the velocities, temperatures, and pressures across a tropical cyclone.

While near the shore, tropical cyclones can cause damage due to storm-surge flooding, wind-wave battering, beach erosion, wind damage, heavy rain, tornadoes, and lightning. The surge is caused by Ekman transport of water toward shore, and by the reduced atmospheric pressure head within the eye.

Science Graffito

In 1989, category 5 hurricane Hugo moved directly over the US. Virgin Islands in the Caribbean Sea. When the hurricane reached the island of St. Croix, it temporarily stopped its westward translation, allowing the intense eye wall to blast the island with violent winds for hours. The following is an eyewitness account.

"It had been many years since St. Croix was in the path of a major storm. Hurricane Hugo reached into the Lesser Antilles with a deliberate vengeance. St. Croix was somewhat prepared. Many hundreds of people had moved into schools and churches to take refuge. But no one was ready for what happened next. By 1800 hours winds were a steady 50 kts [25 m s^{-1}] with gusts up to 70 kts [35 m s^{-1}] from the northwest. I was on the top floor of the wooden Rectory at the St. Patrick's Church in Frederiksted with my husband and 8 month old son."

"By 2000 hours it was apparent that our comfortable room with a view was not going to provide a safe haven. The electricity had been out for some time and a very big gust from the north blew the air conditioner out of the window, landing at the foot of our bed. We evacuated with only one diaper change and bottle of baby juice, leaving behind the playpen, high chair, and bundles of accessories brought from home. We followed Fr. Mike down the wooden staircase. Drafts were everywhere and glass doors exploded just as we passed on our way to Fr. O'Connor's main living quarters on the first floor, where the walls were made of thick coral blocks."

"We settled in again in spite of the persistent crashing and banging against the heavy wooden shutters. We had to shout to hear each other across the room and our ears were popping. In the bathroom, the plumbing sounded like a raging sea. The water in the toilet bowl sloshed around and vibrated. Mercifully the baby slept."

"Soon the thick concrete walls and floors were vibrating, accompanied by a hum that turned into the 'freight train howl'. The banging intensified and persisted for the next 4 hours. By 0100 hours we were tense and sweaty and wondering if it would ever end and if there was anything left outside. Fr. O'Connor was praying and feared that many people must be dead. He got up to open a closet door and a wall of water flowed into the bedroom. At that point we moved to the dining room with a group of 8 other people trapped in the rectory and waited."

"There was concern that the rest of the roof would go and it was decided we would make a run for the schoolhouse made of 2-foot [0.61 m] thick concrete walls. I held the baby in my arms and with flimsy flip-flops [sandals], just about skated across the cement courtyard dodging flying branches and sheets of galvanized aluminum. The window was opened for us as a big, old mahogany tree blocked the door."

"Shortly after, the eye was over us. The thick wooden shutters were flung open and about 100 people outside climbed in the window. The housing project nearby had been stripped of its north and east walls. The eye remained over us for 2 hours then the wind started up with the same intensity coming from the southwest. Only now the room was packed. Strangers were sharing the same mattresses. People slept in desks and chairs made for elementary children, and it was hot. Toddlers and infants wailed."

(continues on next page)

Science Graffito

(continuation)

"There was no generator, only an occasional flashlight could be seen. Fears of [storm] surges were on everyone's' minds. We were only 200 meters from the west shoreline. Hugo had slowed down its eastward track to 4 mph [1.8 m s^{-1}] and the eye passed straight through the middle of the 23-mile [37 km] long island of St. Croix. It seems like the storm was in a fixed permanent position."

"When dawn broke, the winds still howled. By 0800 it was safe to open the windows and the landscape made me burst into tears. There was not a leaf left on a tree, there was not a tree left standing, just tangled branches lying sideways everywhere and not one blade of green grass. The wind had burned the ground and turned everything brown. The gray skies, light rain, and brown landscape persisted for several weeks."

"There were only 2 deaths reported but within weeks several dozen people died from heart attacks, strokes, electrocutions and other accidents associated with reconstruction. The majority of the island residents functioned without power for 3 to 6 months, using generators or candle power and gas stoves."

– Susan Krueger Allick Beach, 1999

16.11. HOMEWORK EXERCISES

16.11.1. Broaden Knowledge & Comprehension

B1. a. Search for satellite images and movies for a few recent tropical cyclones. Discuss similarities and differences of the tropical cyclone appearance.
b. Same, but for radar images during landfall.
c. Same, but for photos from hurricane hunters.

B2. Search for web sites that show tropical cyclone tracks for: a. the current tropical cyclone season.
b. past tropical cyclone seasons.

B3. What names will be used for tropical cyclones in the next tropical cyclone season, for the ocean basin assigned by your instructor?

B4. How many NWP models are available for forecasting tropical cyclone track, & how are they used?

B5. What is the long range forecast for the number of tropical cyclones for the upcoming season, for an ocean basin assigned by your instructor. (Or, if the season is already in progress, how are actual numbers and intensities comparing to the forecast.)

B6. Find maps of "tropical cyclone potential", "tropical cyclone energy", "intensity", or "sea-surface temperature". What are these products based on, and how are they used for tropical cyclone forecasting?

B7. Search the web for photos and info on **hot convective towers**. How do they affect tropical cyclones?

16.11.2. Apply

A1. At 10° latitude, find the absolute angular momentum (m^2 s^{-1}) associated with the following radii and tangential velocities:

	R (km)	M_{tan} (m s^{-1})
a.	50	50
b.	100	30
c.	200	20
d.	500	5
e.	1000	0
f.	30	85
g.	75	40
h.	300	10

A2. If there is no rotation in the air at initial radius 500 km and latitude 10°, find the tangential velocity (m s^{-1} and km h^{-1}) at radii (km):
a. 450 b. 400 c. 350 d. 300
e. 250 f. 200 g. 150 h. 100

A3. Assume ρ = 1 kg m^{-3}, and latitude 20° Find the value of gradient wind (m s^{-1} and km h^{-1}) for:

	R (km)	$\Delta P/\Delta R$ (kPa/100 km)
a.	100	5
b.	75	8
c.	50	10
d.	25	15
e.	100	10
f.	75	10
g.	50	20
h.	25	25

However, if a gradient wind is not possible for those conditions, explain why.

A4. For the previous problem, find the value of cyclostrophic wind (m s^{-1} and km h^{-1}).

A5. Plot pressure vs. radial distance for the max pressure gradient that is admitted by gradient-wind theory at the top of a tropical cyclone for the latitudes (°) listed below. Use z = 17 km, P_o = 8.8 kPa.
a. 5 b. 7 c. 9 d. 11 e. 13 f. 17 g. 19
h. 21 i. 23 j. 25 k. 27 m. 29 n. 31 o. 33

A6. At sea level, the pressure in the eye is 93 kPa and that outside is 100 kPa. Find the corresponding pressure difference (kPa) at the top of the tropical cyclone, assuming that the core (averaged over the tropical cyclone depth) is warmer than surroundings by (°C):
a. 5 b. 2 c. 3 d. 4

e. 1 f. 7 g. 10 h. 15

A7. At radius 50 km the tangential velocity decreases from 35 m s^{-1} at the surface to 10 m s^{-1} at the altitude (km) given below:
a. 2 b. 4 c. 6 d. 8
e. 10 f. 12 g. 14 h. 16
Find the radial temperature gradient (°C/100km) in the tropical cyclone. The latitude = 10°, and average temperature = 0°C.

A8. Find the total entropy ($J \cdot kg^{-1} \cdot K^{-1}$) for:

	P (kPa)	T (°C)	r (g kg^{-1})
a.	100	26	22
b.	100	26	0.9
c.	90	26	24
d.	80	26	0.5
e.	100	30	25
f.	100	30	2.0
g.	90	30	28
h.	20	–36	0.2

A9. On a thermo diagram of the Atmospheric Stability chapter, plot the data points from Table 16-5. Discuss.

A10. Starting with saturated air at sea-level pressure of 90 kPa in the eye wall with temperature of 26°C, calculate (by equation or by thermo diagram) the thermodynamic state of that air parcel as it moves to:
a. 20 kPa moist adiabatically, and thence to
b. a point where the potential temperature is the same as that at 100 kPa at 26°C, but at the same height as in part (a). Thence to
c. 100 kPa dry adiabatically and conserving humidity. Thence to
d. Back to the initial state.
e to h: Same as a to d, but with initial T = 30°C.

A11. Given the data from Table 16-5, what would be the mechanical energy (J) available if the average temperature at the top of the tropical cyclone were
a. –18 b. –25 c. –35 d. –45
e. –55 f. –65 g. –75 h. –83

A12. For the previous problem, find the minimum possible eye pressure (kPa) that could be supported.

A13. Use P_∞ = 100 kPa at the surface. What maximum tangential velocity (m s^{-1} and km h^{-1}) is expected for an eye pressure (kPa) of:
a. 86 b. 88 c. 90 d. 92
e. 94 f. 96 g. 98 h. 100

A14. For the previous problem, what are the peak velocity values (m s^{-1} and km h^{-1}) to the right and left of the storm track, if the tropical cyclone translates with speed (m s^{-1}):
(i) 2 (ii) 4 (iii) 6 (iv) 8 (v) 10
(vi) 12 (vii) 14 (viii) 16 (ix) 18 (x) 20

A15. For radius (km) of:
a. 5 b. 10 c. 15 d. 20
e. 25 f. 30 g. 50 h. 100
find the tropical cyclone-model values of pressure (kPa), temperature (°C), and wind components (m s^{-1}), given a pressure in the eye of 95 kPa, critical radius of R_o = 20 km, and W_s = –0.2 m s^{-1}. Assume the vertically-averaged temperature in the eye is 0°C.

A16. (§) For the previous problem, plot the radial profiles of those variables between radii of 0 to 200 km.

A17. Use P_∞ = 100 kPa at the surface. Find the pressure-head contribution to rise of sea level (m) in the eye of a tropical cyclone with central pressure (kPa) of:
a. 86 b. 88 c. 90 d. 92
e. 94 f. 96 g. 98 h. 100

A18. Find the Ekman transport rate [$km^3/(h \cdot km)$] and surge slope (m km^{-1}) if winds (m s^{-1}) of:
a. 10 b. 20 c. 30 d. 40
e. 50 f. 60 g. 70 h. 80
in advance of a tropical cyclone are blowing parallel to the shore, over an ocean of depth 50 m. Use C_D = 0.005 and assume a latitude of 30°.

A19. What is the Kelvin wave speed (m s^{-1} and km h^{-1}) in an ocean of depth (m):
a. 200 b. 150 c. 100 d. 80
e. 60 f. 40 g. 20 h. 10

A20. For the previous problem, find the growth rate of the Kelvin wave amplitude (m h^{-1}) if the tropical cyclone tracks south parallel to shore at the same speed as the wave.

A21. Find the wind-wave height (m) and wavelength (m) expected for wind speeds (m s^{-1}) of:
a. 10 b. 15 c. 20 d. 25
e. 30 f. 40 g. 50 h. 60

A22. For the previous problem, give the:
(i) Beaufort wind category, and give a modern description of the conditions on land and sea
(ii) Saffir-Simpson tropical cyclone Wind category, its corresponding concise statement, and describe the damage expected.

16.11.3. Evaluate & Analyze

E1. In Fig. 16.2, if the thin ring of darker shading represents heavy precipitation from the eyewall thunderstorms, what can you infer is happening in most of the remainder of the image, where the shading is lighter grey? Justify your inference.

E2. Thunderstorm depths nearly equal their diameters. Explain why tropical cyclone depths are much less than their diameters. (Hint, consider Fig. 16.3)

E3. If you could see movie loops of satellite images for the same storms shown in Figs. 16.1 and 16.4 in the Northern Hemisphere, would you expect these satellite loops to show the tropical cyclone clouds to be rotating clockwise or counterclockwise? Why?

E4. Speculate on why tropical cyclones can have eyes, but mid-latitude supercell thunderstorms do not?

E5. If a tropical cyclone with max sustained wind speed of 40 m s^{-1} contains tornadoes that do EF4 damage, what Saffir-Simpson Wind Scale category would you assign to it? Why?

E6. Why don't tropical cyclones or typhoons hit the Pacific Northwest coast of the USA and Canada?

E7. The INFO box on tropical cyclone-induced Currents in the ocean shows how Ekman transport can lower sea level under a tropical cyclone. However, we usually associate rising sea level with tropical cyclones. Why?

E8. Consider Fig. 16.12. Should there also be a jet along the north edge of the hot Saharan air? If so, explain its characteristics. If not, explain why.

E9. Fig. 16.14a shows wind moving from east to west through an easterly wave. Fig. 16.14b shows the whole wave moving from east to west. Can both these figures be correct? Justify your answer.

E10. In the Extratropical Cyclone chapter, troughs were shown as southward meanders of the polar jet stream. However, in Fig. 16.14 the troughs are shown as northward meanders of the trade winds. Explain this difference. Hint, consider the General Circulation.

E11. Compare and contrast a TUTT with mid-latitude cyclone dynamics as was discussed in the Extratropical Cyclone chapter.

E12. Consider Fig. 16.16. If a typhoon in the Southern Hemisphere was blown by the trade winds toward the Northern Hemisphere, explain what would happen to the tropical cyclone as it approaches the equator, when it is over the equator, and when it reaches the Northern Hemisphere. Justify your reasoning.

E13. Fig. 16.17 suggests that a cold front can help create a tropical cyclone, while Fig. 16.23 suggests that a cold front will destroy a tropical cyclone. Which is correct? Justify your answer. If both are correct, then how would you decide on tropical cyclogenesis or cyclolysis?

E14. In the life cycle of tropical cyclones, Mesoscale Convective Systems (MCSs) are known to be one of the possible initial stages. But MCSs also occur over the USA, as was discussed in the Thunderstorm chapter. Explain why the MCSs over the USA don't become tropical cyclones.

E15. Based on the visible satellite image of Fig. 16.19, compare and contrast the appearance of the tropical cyclone at (c) and the tropical storm at (d). Namely, what clues can you use from satellite images to help you decide if the storm has reached full tropical cyclone strength, or is likely to be only a tropical storm?

E16. What are the other large cloud areas in Fig. 16.19 that were not identified in the caption? Hint, review the Satellites & Radar chapter.

E17. The Bermuda High (or Azores High) as shown in Fig. 16.20 forms because the ocean is cooler than the surrounding continents. However, tropical cyclones form only when the sea-surface temperature is exceptionally warm. Explain this contradiction.

E18. If global warming allowed sea-surface temperatures to be warmer than 26.5°C as far north as 60°N, could Atlantic tropical cyclones reach Europe? Why?

E19. Could a tropical cyclone exist for more than a month? Explain.

E20. a. What is the relationship between angular momentum and vorticity?
b. Re-express eq. (16.1) as a function of vorticity.

E21. From the Atmospheric Forces & Winds chapter, recall the equation for the boundary-layer gradient wind. Does this equation apply to tropical cyclone boundary layers? If so what are it's limitations and characteristics.

E22. The top of Fig. 16.24 shows weak low pressure in the eye surrounding by high pressure in the eyewall at the top of the storm. Explain why the storm cannot have high pressure also in the eye.

E23. Fig. 16.25 shows the wind vector M parallel to the acceleration vector F_{net}. Is that realistic? If not, sketch the likely vectors for M and F_{net}. Explain.

E24. The words "**supergradient winds**" means winds faster that the gradient wind speed. Are the outflow winds at the top of a tropical cyclone supergradient? Justify your answer.

E25.(§) Using relationships from Chapter 1, plot an environmental pressure profile vs. height across the troposphere assuming an average temperature of 273 K. Assume the environmental sea-level pressure is 100 kPa. On the same graph, plot the pressure profile for a warm tropical cyclone core of average temperature (K): a. 280 b. 290 c. 300 assuming a sea-level pressure of 95 kPa. Discuss.

E26. Suppose that the pressure-difference magnitude between the eye and surroundings at the tropical cyclone top is only half that at the surface. How would that change, if at all, the temperature model for the tropical cyclone? Assume the sea-level pressure distribution is unaltered.

E27. Create a table that has a column listing attributes of mid-latitude cyclones, and another column listing attributes of tropical cyclones. Identify similarities and differences.

E28. The circles in Fig. 16.29 illustrate the hypsometric situation applied to the warm core of a tropical cyclone. Why cannot the hypsometric equation explain what drives subsidence in the eye?

E29. In Fig. 16.30b, why does the T_d line from Point 1 to Point 2 follow a contour that slopes upward to the right, even though the air parcel in Fig 16.30a is moving horizontally, staying near sea level?

E30. a. Re-express eq. (16.7) in terms of potential temperature.

b. Discuss the relationship between entropy and potential temperature.

E31. What factors might prevent a tropical cyclone from being perfectly efficient at extracting the maximum possible mechanical energy for any given thermodynamic state?

E32. For the tropical cyclone model given in this chapter, describe how the pressure, tangential velocity, radial velocity, vertical velocity, and temperature distribution are consistent with each other, based on dynamic and thermodynamic relationships. If they are not consistent, quantify the source and magnitude of the discrepancy, and discuss the implications and limitations. Consider the idealizations of the figure below. (ABL = atmos. boundary layer.)

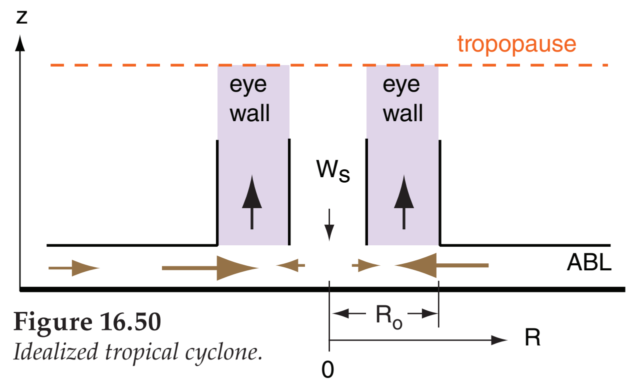


Figure 16.50
Idealized tropical cyclone.

Hints: a. Use the cyclostrophic relationship to show that tangential velocity is consistent with the pressure distribution.

b. Use mass conservation for inflowing air trapped within the boundary layer to show how radial velocity should change with R, for $R > R_o$.

c. Assuming wave drag causes radial velocity to be proportional to tangential velocity squared, show that the radial velocity and tangential velocity equations are consistent. For $R < R_o$, use the following alternate relationship based on observations in tropical cyclones: $M_{tan}/M_{max} = (R/R_o)^2$.

d. Rising air entering the bottom of the eye wall comes from two sources, the radial inflow in the boundary layer from $R > R_o$, and from the subsiding air in the eye, which hits the ground and is forced to diverge horizontally in order to conserve mass. Combine these two sources of air to compute the average updraft velocity within the eye wall.

e. Use mass continuity in cylindrical coordinates to derive vertical velocity from radial velocity, for $R < R_o$. (Note, for $R > R_o$, it was already assumed in part (a) that air is trapped in the ABL, so there is zero vertical velocity there.)

f. Use the hypsometric relationship, along with the simplifications described in the temperature subsection of the tropical cyclone model, to relate the radial temperature distribution (averaged over the whole tropical cyclone depth) to the pressure distribution.

E33. An article by Willoughby and Black (1996: tropical cyclone Andrew in Florida: dynamics of a disaster, *Bull. Amer. Meteor. Soc.*, **77**, 543-549) shows tangential wind speed vs. radial distance.

a. For their Figs. 3b and 3d, compare their observations with the tropical cyclone model in this chapter.

b. For their Figs. 3e - 3g, determine translation speed of the tropical cyclone and how it varied with time as the storm struck Florida.

E34. To reduce fatalities from tropical cyclones, argue the pros and cons of employing better mitigation technology and the pros and cons of population control. Hint: Consider the whole world, including issues such as carrying capacity (i.e., the finiteness of natural resources and energy), sustainability, politics and culture.

E35. In Fig. 16.41, one relative maximum (A) of storm-surge height is directly in front of the storm's path, while the other (B) is at the right front quadrant. Explain what effects could cause each of these surges, and explain which one might dominate further from shore while the other might dominate when the tropical cyclone is closer to shore.

E36. Fig. 16.43 for a storm surge caused by Ekman transport is for a tropical cyclone just off the east coast of a continent. Instead, suppose there was a tropical cyclone-force cyclone just off the west coast of a continent at mid-latitudes.

a. Would there still be a storm surge caused by Ekman transport?

b. Which way would the resulting Kelvin wave move (north or south) along the west coast?

E37. Devise a mathematical relationship between the Saffir-Simpson tropical cyclone Wind scale, and the Beaufort wind scale.

E38. Estimate CDP using the data in Fig. 16.34.

16.11.4. Synthesize

S1. What if the Earth rotated twice as fast. Describe changes to tropical cyclone characteristics, if any.

S2. What if the average number of tropical cyclones tripled. How would the momentum, heat, and moisture transport by tropical cyclones change the global circulation, if at all?

S3. Some science fiction novels describe "supercanes" with supersonic wind speeds. Are these physically possible? Describe the dynamics and thermodynamics necessary to support such a storm in steady state, or use the same physics to show why they are not possible.

S4. Suppose the tropical tropopause was at 8 km altitude, instead of roughly 16 km altitude. How would tropical cyclone characteristics change, if at all?

S5. What if the Earth's climate was such that the tropics were cold and the poles were hot, with sea surface temperature greater than 26°C reaching from the poles to 60° latitude. Describe changes to tropical cyclone characteristics, if any.

S6. Suppose that the sea surface was perfectly smooth, regardless of the wind speed. How would tropical cyclone characteristics change, if at all?

S7. Is it possible to have a tropical cyclone without a warm core? Be aware that in the real atmosphere, there are tropical cyclone-force cyclones a couple times a year over the northern Pacific Ocean, during winter.

S8. Suppose that the thermodynamics of tropical cyclones were such that air parcels, upon reaching the top of the eye wall clouds, do not loose any heat by IR cooling as they horizontally diverge away from the top of the tropical cyclone. How would the Carnot cycle change, if at all, and how would that affect tropical cyclone intensity?

S9. Is it possible for two tropical cyclones to merge into one? If so, explain the dynamics and thermodynamics involved. If so, is it likely that this could happen? Why?

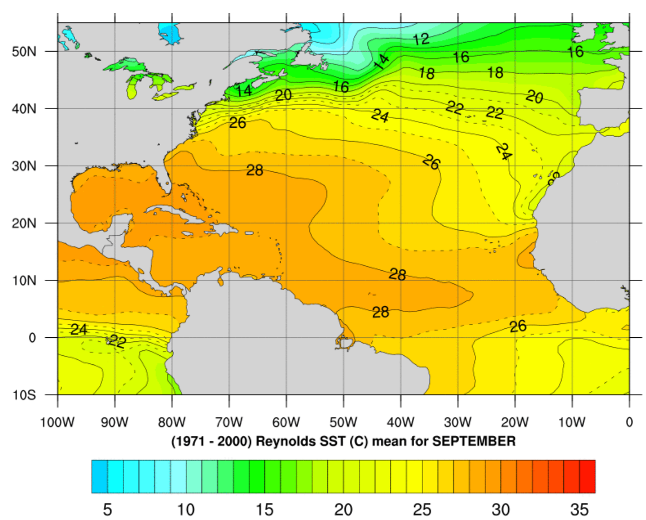


Figure 16.51
Sea surface temperature (°C) in the Atlantic Ocean averaged over 30 years of Septembers. Orange regions favor hurricane formation. [Courtesy of NWS National Hurricane Ctr. image.]