# ESCI 344 – Tropical Meteorology
## Lesson 10 – Tropical Cyclones: Formation and Structure

**References:** *A Global View of Tropical Cyclones*, Elsberry (ed.)
*The Hurricane*, Pielke
*Tropical Cyclones: Their evolution, structure, and effects*, Anthes
*Forecasters' Guide to Tropical Meteorology*, Atkinsson
*Forecasters Guide to Tropical Meteorology* (updated), Ramage
*Global Guide to Tropical Cyclone Forecasting*, Holland (ed.), online at
*http://www.bom.gov.au/bmrc/pubs/tcguide/globa_guide_intro.htm*

**Reading:** *A Global View of Tropical Cyclones*, Chapter 3, Frank (e-reserve)
*Tropical Cyclones: Their Evolution, Structure, and Effects*, Chapter 2 (e-reserve)
*Hurricane,* Chapter 2, Pielke (e-reserve)
*Global Guide*, Chapter 2 (online)

### GENERAL

- **Tropical cyclones are primarily driven by latent heating.**
- **As the air spirals in toward the center, it picks up latent heat and sensible heat through evaporation from the ocean.**
    - **Once the air is saturated, it can still pick up some sensible heat, but latent heating is the dominant mechanism.**
- **As the air approaches the center of the vortex, it rises in convection either in the eye wall, or in the spiral convective bands. As it rises, it cools and the water vapor condenses, giving off its latent heat.**
- **The warming of the air due to the latent heat release results in low-level height falls, and upper-level height rises, which helps maintain the low-level convergence of the warm-moist air.**
- **The air is exhausted out and away at the upper levels.**

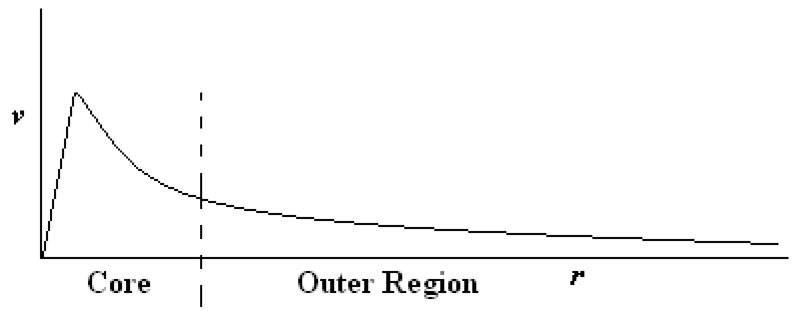
### STRENGTH, SIZE, AND INTENSITY

- **We will use the following definitions:**
    - ***Core intensity*, based on maximum winds or minimum sea-level pressure.**
    - ***Size*, based on the mean radius of the outermost closed isobar.**
    - ***Strength*, based on the shape of the outer core wind profile.**
- **There is great variability in the size, intensity, and strength of tropical cyclones.**
    - **Storms can be large and intense, small and intense, large and weak, etc.**

- ○ **You can't infer intensity based on size or strength.**

**DISTRIBUTION OF WIND AND ANGULAR MOMENTUM**

- **The diagram below shows the typical tangential wind structure in a tropical cyclone.**




- **The wind structure is often represented by a modified Rankin vortex,**

$$vr^{a} = C \quad r > r_{\max\, wind}$$
$$vr^{-1} = D \quad r > r_{\max\, wind}$$

**where *C*, *D*, and *a* are empirically determined constants.**

  - ○ ***a* is usually 0.4 to 0.6.**
  - ○ ***a* = 1 would be a pure Rankin vortex, in which relative angular momentum is conserved.**

- **In cylindrical coordinates, the radial and tangential components of the momentum equation are**

$$\frac{Du}{Dt} - fv - \frac{v^2}{r} = -\frac{1}{\rho}\frac{\partial p}{\partial r} + F_r$$
$$\frac{Dv}{Dt} + fu - \frac{uv}{r} = -\frac{1}{\rho r}\frac{\partial p}{\partial \theta} + F_\theta$$

**where $u$ is the radial velocity, $v$ is the tangential velocity, $r$ is the distance from the center of the storm, $\theta$ is the angular measure, and $F_r$ and $F_\theta$ represent turbulent friction.**

- **If the vortex is steady, axisymmetric, and friction is ignored, then the tangential wind is**

$$\frac{v^2}{r} + fv - \frac{1}{\rho}\frac{\partial p}{\partial r} = 0\,, \qquad \textbf{(1)}$$

**so the vortex is in gradient balance.**

- **Dividing by *fv*, we get**

$$\frac{v}{f\,r} + 1 - \frac{1}{f\,v\rho}\frac{\partial p}{\partial r} = 0\,.$$

- **The first term is just the Rossby number, $R_O$.**
  - **In the core of the vortex, the Rossby number is large, so the Coriolis effects can be ignored. Therefore, the core of the storm is in cyclostrophic balance.**
  - **Outside the core, the Rossby number is of the order of unity, so gradient balance holds.**
- **In either case, the wind speed depends on the pressure gradient.**
- **Broadly speaking, the lower the central pressure, the faster the maximum wind will be.**
  - **The relation between maximum winds and the central pressure is closely approximated by**

$$v_{\max} = A\sqrt{p_\infty - p_c}$$

    **where $p_\infty$ is the ambient sea-level pressure outside of the circulation, and $p_c$ is the minimum central pressure.**
  - **The constant *A* is empirically determined. A value of 6.3 is often used for Atlantic hurricanes.**
- **If no energy were added or subtracted from an air parcel as it spiraled toward the center of the vortex, then its absolute angular momentum would have to be conserved.**
  - **The absolute angular momentum is the relative angular momentum plus the angular momentum due to the rotation of the Earth.**
  - **A ring of air parcels surrounding the center point of the vortex at distance *r* and stationary with respect to the Earth will have a specific angular momentum (angular momentum per unit mass) of**

$$M_f = \omega r^2 = f_0 r^2/2$$

    **just due to the rotation of the Earth.**

- **If the parcels have a tangential velocity, *v*, then their specific relative angular momentum is**

$$M = vr\,,$$

**so the specific absolute angular momentum is**

$$M_a = vr + f_0 r^2/2\,. \qquad \textbf{(2)}$$

- **If the parcel started out at rest at a distance of 500 km, and spiraled in to a radius of 15 km, it would have attained a tangential velocity of over 600 m/s if its angular momentum were conserved.**

- **Obviously, parcels don't conserve angular momentum as they spiral into the center of a tropical cyclone.**
- **In fact, the absolute angular momentum decreases toward the center of tropical cyclones, which means that air parcels must be losing angular momentum as they spiral inward.**
  - **The parcels lose angular momentum through turbulent dissipation.**
- **The absolute vorticity of an axisymmetric vortex is**

$$\eta = f_0 + \frac{\partial v}{\partial r} + \frac{v}{r}\,, \qquad \textbf{(3)}$$

**where the second and third terms are just the shear and curvature terms. From equation (2) we can show that the vorticity and the specific absolute angular momentum are related via**

$$\eta = \frac{1}{r}\frac{\partial M_a}{\partial r}\,. \qquad \textbf{(4)}$$

## INERTIAL STABILITY OF A VORTEX

- **A fundamental parameter for assessing how a vortex interacts with its environment is the inertial stability, which is developed mathematically below.**
- **The radial momentum equation in an axisymmetric vortex without friction is**

$$\frac{Du}{Dt} = f_0 v + \frac{v^2}{r} - \frac{1}{\rho}\frac{\partial p}{\partial r}\,.$$

- **In terms of specific absolute angular momentum this can be written as**

$$\frac{Du}{Dt} = \frac{M_a^2}{r^3} - \frac{1}{r^3}\left(\frac{f_0^2 r^4}{2} + \frac{r^3}{\rho}\frac{\partial p}{\partial r}\right)$$

**or**

$$\frac{Du}{Dt} = \frac{1}{r^3}\left[M_a^2 - G^2(r)\right]. \tag{5}$$

**where**

$$G^2(r) = \frac{f_0^2 r^4}{2} + \frac{r^3}{\rho}\frac{\partial p}{\partial r}. \tag{6}$$

- **Notice that for an air parcel that is in gradient balance, $G(r) = M_a(r)$.**
- **You can think of $G(r)$ as being the angular momentum that an air parcel in gradient balance would have at radius $r$.**

- **Now, imagine that an air parcel at position $r_0$ is initially in gradient balance with its surroundings. In this case its absolute angular momentum, $M_0$, will be**

$$M_0 = G(r_0). \tag{7}$$

  - **If the parcel is displaced radially a small distance, $\delta r$, its absolute angular momentum will be conserved and will remain equal to $M_0$ as it is displaced. However, $G(r)$ can be represented by the first few terms of a Taylor series expansion around $r$ as**

$$G^2(r) \cong G^2(r_0) + \frac{\partial G^2}{\partial r}\delta r. \tag{8}$$

  **Using equations (7) and (8) in equation (5) we get that for the displaced air parcel**

$$\frac{D^2(\delta r)}{Dt^2} = -\frac{1}{r^3}\frac{\partial G^2}{\partial r}\delta r. \tag{9}$$

  - **We know that $G(r)$ is the absolute angular momentum that an air parcel in gradient balance would have at radius $r$, so equation (9) becomes**

$$\frac{D^2(\delta r)}{Dt^2} + \frac{1}{r^3}\frac{\partial M_a^2}{\partial r}\delta r = 0. \tag{10}$$

- **Solutions to equation (10) are oscillations with frequency of**

$$\omega^2 = \frac{1}{r^3}\frac{\partial M_a^2}{\partial r}. \tag{11}$$

- **If the frequency is real, then the parcel will just oscillate around its original radius, and the flow is *inertially stable*. If the frequency is imaginary, then the**

**parcel will accelerate away from its original radius, and the flow is *inertially* unstable.**

- **The conditions for inertial stability are**

$$\omega^2 > 0 \quad \text{Inertially stable}$$
$$\omega^2 = 0 \quad \text{Inertially neutral}$$
$$\omega^2 < 0 \quad \text{Inertially unstable}$$

- **Since absolute angular momentum and vorticity are related, we can write the oscillation frequency in terms of vorticity, as**

$$\omega^2 = \eta\left(f_0 + 2v/r\right). \quad \textbf{(12)}$$

- **The more inertially stable a vortex is, the less it will interact horizontally with its environment, since horizontal displacements are resisted. We will apply this concept when discussing the structure of the core and outer regions of a tropical cyclone.**
- **Physically, inertial stability can be explained as follows:**
  - **If the parcel is displaced outward, and in its new position find that the Coriolis force is stronger than the new pressure gradient force, then it will accelerate outward away from its initial position and the vortex is inertially unstable.**
  - **If the parcel is displaced outward, and in its new position find that the Coriolis force is weaker than the new pressure gradient force, then it will accelerate inward toward its initial position and the vortex is inertially stable.**
- **In straight-line flow, or for weak vortexes where the curvature term in equation (12) can be ignored, equation (12) becomes**

$$\omega^2 = \eta f_0. \quad \textbf{(13)}$$

  - **This is why you often hear it said that anytime absolute vorticity is negative that the flow is inertially unstable. However, keep in mind that *<u>for stronger vortexes, particularly at low latitudes, negative absolute vorticity doesn't automatically imply inertial instability</u>*. You must look at the radial gradient of absolute angular momentum to assess the inertial stability in these cases.**

## ROSSBY RADIUS OF DEFORMATION

- **A fundamental horizontal length scale for a disturbance in a rotating fluid is the *Rossby radius of deformation*.**
- **The Rossby radius of deformation is the distance that a gravity wave (which are the means by which the fluid adjusts to equilibrium) will travel in one inertial period ($\varpi^{-1}$).**
- **The Rossby radius of deformation is therefore**

$$\lambda_R = \frac{c_g}{\sqrt{\eta(f_0 + 2v/r)}} \qquad \textbf{(14)}$$

**where $c$ is the horizontal group velocity a gravity wave, and the denominator is the inertial frequency from equation (12).**

- ○ **Note: For flows whose absolute vorticity is primarily due to planetary vorticity (i.e., flows where $\zeta << f$), equation (14) becomes**

$$\lambda_R = c_g / f_0 ,$$

**which is the form of the Rossby radius of deformation most commonly used in dynamic meteorology textbooks. However, keep in mind that it is just a specific form of the more general definition in equation (14).**

- **The gravity wave group velocity, $c_g$, depends on the stratification of the fluid.**
  - ○ **For a barotropic fluid, it is given by $c_g = \sqrt{gH}$ where $H$ is the depth of the fluid.**
  - ○ **For a baroclinic fluid, there are multiple *baroclinic modes* of oscillation, each with its own group velocity. For waves that travel primarily horizontally (long wavelengths) the group velocity can be approximated as**

$$c_g \cong NH/n\pi$$

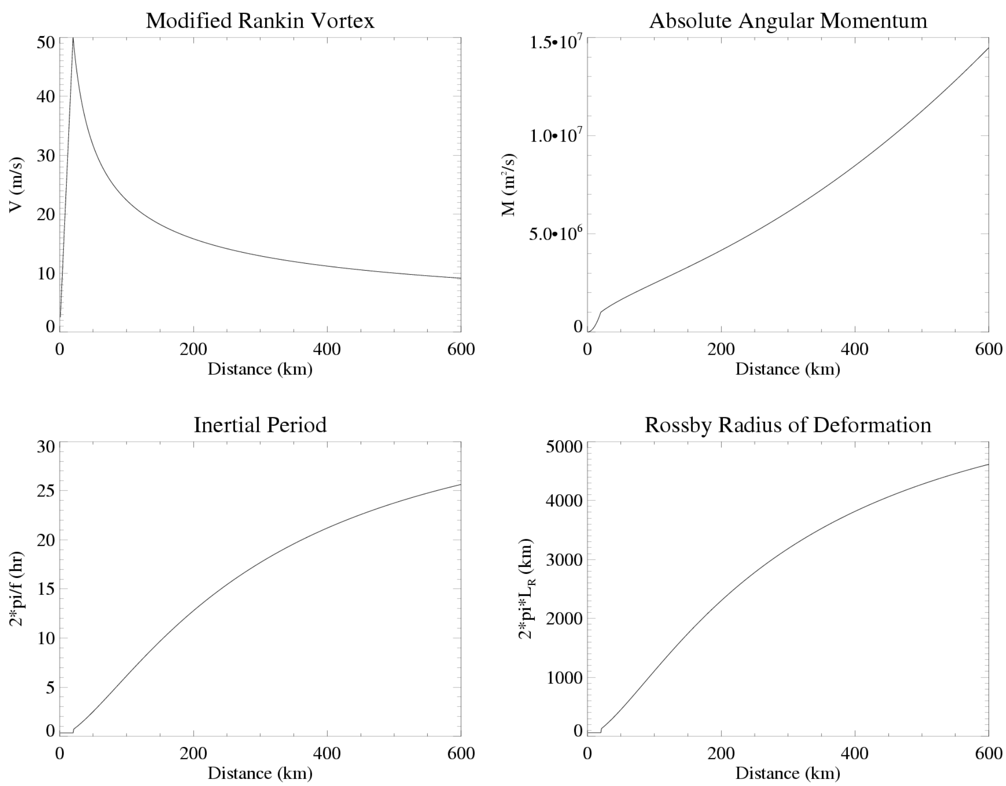
**where $N$ is the Brunt-Vaisala frequency, $H$ is the vertical scale of the disturbance, and $n$ refers to the different baroclinic modes of oscillation ($n$ = 1, 2, 3, …).**

  - ○ **The baroclinic modes have a significantly slower group velocity than does the barotropic mode, and so the Rossby radius of deformation is smaller for the baroclinic modes.**

- **The response of the fluid to a disturbance can be assessed by comparing the horizontal length scale of the disturbance, *L*, to the Rossby radius of deformation.**
  - **For disturbances whose horizontal length scale, *L*, is much less than $2\pi\lambda_R$ , the mass field will adjust to the velocity field.**
  - **For disturbances whose horizontal length scale, *L*, is much greater than $2\pi\lambda_R$, the mass field will adjust to the velocity field.**
  - **For disturbances in the intermediate range, mutual adjustment occurs.**
- **The size of the disturbance compared to the Rossby radius of deformation can also give us an idea of how persistent a circulation can be.**
  - **For disturbances whose size is comparable or larger than the Rossby radius of deformation, we know that the wind will adjust to the mass field, while for small disturbances the mass will adjust to the wind field.**
  - **Diabatic heating will cause the mass field to be perturbed.**
  - **If the heating is confined to a region much smaller than the Rossby radius of deformation, the disturbance in the mass field will not influence the wind.**
    - **All that will happen is that gravity waves will propagate away from the disturbance, and eventually nothing will be left of it.**
  - **Only if the heating is over a region comparable to the Rossby radius of deformation will a circulation in the wind field develop in response.**
- **Disturbances whose sizes are much less than the Rossby radius of deformation tend to die out quickly, and not develop a persistent circulation.**

## STRUCTURE AND DYNAMICS OF A TROPICAL CYCLONE

- **The diagrams below show the wind speed, angular momentum, inertial period, and Rossby radius of deformation (*c* = 50 m/s) for a modified Rankin vortex at a latitude of 20°, with a radius of maximum winds of 20 km, and an intensity of 50 m/s (*a* = 0.5).**




  - **In particular, note that the Rossby radius of deformation is very large in the outer region of the cyclone, whereas it is much smaller in the core.**
  - **Also notice the difference in the inertial period, which means the inertial stability is very different in the core region versus the outer region.**

## STRUCTURE OF THE CORE REGION

- **The core region is 3 to 6 times the radius of maximum winds.**
- **The Rossby number (not shown in graphs above) is large in the core region, so the balance is cyclostrophic.**
- **The wind structure of the core has the following general characteristics:**
  - **Radius of maximum winds is generally in the eye wall.**
  - **There is a sharp increase in speed near the radius of maximum winds.**
  - **Winds decrease aloft.**

- **Maximum winds tilt outward with height.**
- **Maximum winds are found in the front-right quadrant of the cyclone.**
  - **This is true, *even if the forward speed of the storm is factored out*.**
  - **The reason isn't completely understood.**

- **Strongest horizontal temperature gradients are near the radius of maximum winds.**
- **Warm core is narrow in the lower troposphere, and bulges outward aloft.**
- **Some features of the core region are the eye and rainbands.**
- **Some definitions:**
  - ***Rainband* – Also called *spiral bands*, or *feeder bands*, they are areas of cloud and of precipitation spiraling in toward the center of the circulation.**
  - ***Convective ring* – Rainbands that near completely circle the center.**
  - ***Moat* – Clear region between two convective rings.**
  - ***Eyewall* – Innermost convective ring.**
  - ***Eye* – Clear region at center of circulation, surrounded by the eyewall.**

## DYNAMICS OF THE CORE REGION

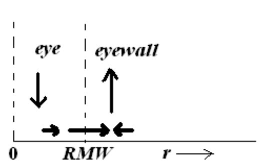
- **The inner core region has very high inertial stability due to the large absolute vorticity and small radius (refer to equation 12).**
  - **This means it doesn't readily interact with its environment, and is why the inner core tends to be very symmetric.**
- **The inner core regions also has a small Rossby radius of deformation.**
  - **In the core, the wind tends to adjust to changes in the mass field. Any changes in the mass field due to heating/cooling or convergence/divergence will result in changes to the wind.**
  - **The intensity of the cyclone can therefore react rapidly to fluctuations in diabatic and latent heating, as well as fluctuations in vertically integrated divergence.**
- **The dynamics and fluctuations in the core are dominated by primarily by convection and heating.**

**THE EYE**

- **An eye generally doesn't form until the cyclone reaches hurricane or typhoon intensity.**
- **The mechanisms by which an eye is formed and maintained are not completely known. One conceptual model is presented below.**
  - **As air spirals inward toward the center of the cyclone, its speed will increase due to conservation of angular momentum.**
    - **However, if angular momentum were truly conserved, wind speeds of 600 m/s or more would be reached.**
    - **Therefore, angular momentum must be dissipated via convection, turbulence, and other processes that retard the flow.**
  - **The angular momentum balance at a point is given by**

$$\frac{\partial M_a}{\partial t} = -u\frac{\partial M_a}{\partial r} - w\frac{\partial M_a}{\partial z} - F_r\,. \qquad \textbf{(12)}$$

    - **The three terms on the right-hand-side are radial advection, vertical advection, and dissipation of angular momentum.**
  - **At some distance near the center of the vortex, the dissipation of angular momentum can no longer keep up with the horizontal advection. This leads to a horizontal convergence of angular momentum.**
  - **This horizontal convergence of angular momentum leads to an increase in the winds above the value that the pressure gradient can support (i.e., they are *super-gradient*).**
  - **The super-gradient winds develop a radially outward component, since the pressure gradient cannot supply the required centripetal acceleration. The consequences of this are (see figure):**
    - **There is radial convergence at distances outside of the radius of maximum winds.**
    - **There is radial divergence at distances inside the radius of maximum winds.**



- The convergence leads to upward motion, and results in the convection within the eye wall.
  - The resultant convection within the eyewall serves to balance the horizontal convergence of angular momentum through vertical advection.
- The upward motion in the eyewall results in outflow aloft, and high perturbation pressure over the center of the storm.
- The upper-level pressure perturbation, combined with the low-level divergence within the eye itself, results in compensating subsidence in the eye.
- Due to the strong inertial stability in the eyewall, the strongest compensating subsidence is found on the inside of the radius of maximum winds.
- This results in an abrupt clearing just inside the eyewall.

- **As the air subsides, the resultant compressional warming actually works against subsidence (through buoyancy).**
- **All that is needed in order to keep the eye relatively clear is enough subsidence to balance the buoyancy.**
  - **Thus, in the steady state, there doesn't have to be strong subsidence in the eye in order to maintain the eye.**
- **The maintenance of the eyewall is a balance between the horizontal advection of angular momentum and the vertical advection. The radius of the eye can expand or contract depending on this balance.**

## EYEWALL REPLACEMENT CYCLE

- **Sometimes an outer convective ring will form and will often cause the inner ring (the eyewall) to dissipate for two reasons.**
  - **The subsidence from the outer ring suppresses convection in the inner ring.**
  - **The outer ring robs the inner ring of inflow.**
- **Eyewall replacement usually results in a lowering of the intensity, since the new eyewall is at a larger radius, and from angular momentum arguments, would have a lower wind speed.**
  - **Intensity may build again one the new eyewall contracts.**
- **The causes and dynamics of eyewall replacement cycles are not well known.**

- **Eyewall replacement tends to occur in very intense cyclones.**

**RAINBANDS**

- **Rainbands are bands of convection that occur in the core region, and also can extend into the outer region of the cyclone.**
- **Rainbands can be 10's of kilometers wide, and 100's of kilometers in length.**
- **Convection often forms on the inside of the rainbands and moves up and forward through them.**
- **The region between the rainbands is characterized by subsidence.**
- **Rainbands may be of three types:**
  - *Moving spiral bands* **– These bands appear to rotate with the circulation.**
  - *Convective rings* **– A rainband that completely encloses the cyclone.**
    - **A convective ring may transition into an eyewall**
  - *Principle spiral bands* **– A non-moving rainband that wraps into the cyclone on the east side of the cyclone, and feeds lower latitude air into the vortex.**
    - **In a rough sense, principle spiral bands separate the inner region of the cyclone from the outer regions.**
    - **In the inner region, air makes several circuits as it spirals into the core, where in the outer region the air doesn't make it to the core.**

**STRUCTURE OF OUTER REGION**

- **In the outer region, the absolute vorticity is weaker and the radius is larger, so this region has much lower inertial stability than does the core region.**
  - **This means the outer region is less symmetric, and also more readily influenced by the environmental flow.**
- **The outer region also has a larger Rossby radius of deformation than does the core.**
  - **In the outer region the mass field adjusts to the wind field.**
- **The tangential circulation can extend 1000 – 2000 km from the center of the storm.**
- **The radial circulation is much smaller in scale (~600 km).**

- **There is usually subsidence in the outer region, with areas of weak upward motion.**
- **A typical ring-like area of subsidence is found around 5° latitude or so from the center, and is sometimes referred to as the *moat*, since it results in clearing.**

**OUTFLOW REGION**

- **The intense updrafts in the eyewall carry cyclonic vorticity upwards. This results in the upper-level outflow near the cyclone center having cyclonic rotation.**
- **As it spreads outward, it loses positive vorticity and gains negative vorticity, so that it becomes anticyclonic as it moves away.**
- **The outflow region has weak inertial stability, and so doesn't resist horizontal flow.**
    - **The outflow region interacts readily with the environmental flow.**
- **The outflow mainly takes place in one or two *outflow jets* or *channels*.**
    - **These outflow channels are shallow, and are in the upper troposphere.**
- **Cyclones with two outflow channels tend to be more intense than those with a single outflow channel.**

**DIURNAL CYCLES**

- **Tropical cyclone convection shows a pronounced diurnal cycle.**
- **Convection is enhanced in the early morning hours (0300 to 0600 local), likely due to the same reasons that tropical convection over the open oceans has an early morning maximum.**
- **Areal coverage in cirrus clouds is maximum in the late afternoon (around 1800 local time).**

**SYNOPTIC SCALE INFLUENCES**

- **Since tropical cyclone formation requires a pre-existing cyclonic disturbance, they normally form in**
    - **Monsoon troughs**
    - **Tropical waves**

- **Old frontal zones or shear lines**

- **Any enhancement of vorticity is favorable for formation.**
  - **A cold surge in the wintertime hemisphere often will enhance the Equatorial westerlies, and is favorable for formation.**
- **Existing storms can influence the formation and development of new storms.**
  - **For storms that have not yet recurved, the path ahead of the storm is unfavorable for formation of another tropical cyclone, while the wake area behind it is favorable.**
    - **This is due to the large scale vertical motion pattern forced by the tropical cyclone.**
  - **The outflow from an existing storm can also sometimes provide too much shear over a region and suppress formation and development.**
- **Formation of upper-tropospheric outflow jets is key for development, since the upper-level mass divergence must be larger than the lower-level mass convergence.**
- **A north-east quadrant outflow jet is enhanced by linking with**
  - **the subtropical jet**
  - **a tropical upper-tropospheric trough (TUTT)**
  - **a deep mid-latitude trough**
  - **an upper-level cold low**
- **The position of the cyclone in relation to the above features is crucial for an enhanced north-east outflow jet. Although they enhance outflow, which aids deepening of the surface low, if they are too close then the increased westerly shear can inhibit growth.**
- **A south-west quadrant outflow jet is enhanced by an intense upper-level anticyclone in the opposite hemisphere.**

## FORMATION AND INTENSIFICATION OF TROPICAL CYCLONES FROM INITIAL DISTURBANCES

- **Tropical convection acts as a heat engine, taking warm moist air from the surface and converting the latent heat into kinetic energy in the updraft, which is then exhausted into the upper troposphere.**

- **The mechanisms and requirement to get a tropical cloud cluster to develop a sustained circulation and develop into a tropical cyclone are complex, and much is still unknown.**
- **In order for a convective cloud cluster to result in pressure falls at the surface, there must be a net removal of mass from the air column (net vertically integrated divergence).**
  - **Since there is compensating subsidence nearby, outside of a typical convective cloud, there really isn't much integrated mass divergence.**
  - **Pressure really won't fall unless there is a mechanism to remove the mass that is exhausted well away from the convection.**
  - **Compensating subsidence near the convection also serves to decrease the buoyancy within the clouds, because the subsiding air will also warm. This reduces the temperature difference between the in-cloud and outside air.**
  - **This illustrates the importance of upper-level outflow in tropical cyclones.**
- **In addition, even if there are pressure falls at the surface, the atmosphere could just generate gravity waves and readjust back to the original pressure.**
  - **The atmospheric response to the heating is going to depend on how the horizontal scale of the heating compares to the appropriate Rossby radius of deformation.**
- **Once a tropical cyclone forms and has an organized circulation, there are feedback mechanisms that serve to maintain or strengthen the circulation.**
- **Once feedback mechanism is *conditional instability of the second kind* (*CISK*), which is discussed further in the next section.**
- **Another feedback mechanism is the coupling between the strong winds in the boundary layer and the transfer of latent and sensible heat into the air as it spirals into the cyclone.**
  - **Frictional convergence of the tangential winds plays a key role in transporting latent heat into the core of the vortex.**
  - **The stronger the winds, the greater the frictional convergence into the core.**
  - **Also, the stronger the winds, the greater the transfer of latent and sensible heat from the sea surface into the inflowing air.**

- **The formation of an anticyclone aloft by the heating aids in the upper-level mass divergence which is necessary to sustain and intensify the cyclone.**
- **Tropical cyclones can maintain themselves as long as there is sufficient inflow of warm, moist air into the cyclone, and there is adequate outflow aloft. Factors that can cause a tropical cyclone to fluctuate in intensity (either up or down) are:**
  - **Variations in SST.**
  - **Interaction with land, which can result in less evaporation and latent heat inflow.**
    - **Tropical cyclones can momentarily increase in intensity as they make landfall due to enhanced low-level convergence due to the increased friction over land.**
  - **Enhanced or suppressed outflow.**
  - **Increased vertical shear.**
    - **Vertical shear can cause the upper-level anticyclone and outflow to decouple from the low-level inflow. If the outflow is weakened, then the mass that is ejected into the upper troposphere by the convection can subside in the vicinity of the tropical cyclone and weaken the convection, as well as limit the surface pressure falls.**

## CONDITIONAL INSTABILITY OF THE SECOND KIND

- **Tropical cyclones often form from convective cloud clusters.**
- **The tropics are full of convective cloud clusters, yet few of them develop into tropical cyclones.**
- **One theory for how tropical cyclones form and intensify is that of *conditional instability of the second kind* or *CISK*.**
- **In its simplest form, CISK can be explained as follows:**
  - **Latent heating of the atmosphere leads (through the hypsometric relationship) to a lowering of surface pressure.**
  - **The lowering of the surface pressure leads to enhanced radial inflow and convergence, which enhances the convection and latent heat release, which further decreases the surface pressures.**
- **CISK is then a positive feedback loop.**

- **There are several difficulties with using CISK to explain tropical cyclone formation.**
- **One problem is that in order for a circulation to form from the latent heating, the scale of the heating must be of the order of the Rossby radius of deformation ($L \sim 2\pi\lambda_R$).**
  - **In the Tropics, $\lambda_R$ is large even for the baroclinic modes, so the mass field disturbances caused by latent heating from cloud clusters isn't large enough to form circulation.**
- **In order to excite modes with a small Rossby radius of deformation, a heating profile that is concentrated in the lower levels is ideal.**
  - **However, the latent heating profile most tropical cloud clusters is concentrated in the upper troposphere, so it doesn't excite the modes with low Rossby radius of deformation.**
- **Most studies of CISK have involved a linear approximation in which the different wavelengths, or modes, do not interact and exchange energy.**
  - **It is also possible that the non-linear interaction of the wave modes is important, which would allow energy that is input into one mode to be transferred into other modes that do have a horizontal scale closer to their Rossby radii of deformation.**
- **Although CISK has trouble explaining how cloud clusters form into a tropical cyclone, it can be used to explain how a tropical cyclone, once formed, intensifies.**
  - **Once an organized, persistent circulation exists, the Rossby radius of deformation will be smaller due to the effects that absolute vorticity has on the inertial period (see equations 10 and 11). Therefore, the mass disturbance caused by the latent heat can approach the Rossby radius of deformation, and the wind field can respond to the mass disturbance.**