Chapter Four

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Notice

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4. Tropical Cyclone Intensity, Structure, and Structure Change

4.1 Introduction[1]

The evolution of a tropical cyclone through its lifecycle from genesis to decay provides constant forecasting challenges. In this chapter, we review the key drivers and limitations on tropical cyclogenesis; processes leading to tropical cyclone intensity and structure changes (including tropical cyclone intensity classification conventions around the world); and extratropical transition (ET) of a tropical cyclone.

Tropical cyclogenesis

Operationally, tropical cyclogenesis is said to have occurred when mean[2] surface winds in excess of tropical storm force (17 m s^{-1}; 33 knots) are observed. Implicit in this operational genesis criterion is the expectation that the tropical storm has become self-sustaining and can continue to intensify without help from its environment.

Since only around 90 tropical cyclones are observed annually around the globe, it is clear that special conditions are required for tropical cyclogenesis. In the remainder of this section, we will review the necessary conditions for tropical cyclogenesis and the sources of the incipient disturbances.

[1] Much of the material in this chapter is drawn from the chapter on tropical cyclones in *An Introduction to Tropical Meteorology*, a COMET/UCAR online textbook co-authored by myself (lead author Dr. Arlene Laing). The complete textbook is available at

http://www.meted.ucar.edu/tropical/textbook/index.htm.

2Averaging time for surface winds varies by region; details of the different regional averaging times and intensity classifications are given below. "Surface" is defined as 10 meters above the ground.

4.2 Necessary ingredients for tropical cyclogenesis

Necessary ingredients for tropical cyclogenesis

Gray (1968) identified six features of the large-scale tropical environment that were necessary ingredients for tropical cyclogenesis:

1. *sufficient ocean thermal energy [SST > 26°C to a depth of 60 m],*
2. *moist mid-troposphere [measured by 700 hPa relative humidity],*
3. *conditionally unstable atmosphere to support deep convection,*
4. *a maximum in lower troposphere relative vorticity,*
5. *weak vertical shear of the horizontal winds at the genesis site, and*
6. *location at least 5 latitude away from the equator.*

Conditions *(i)-(iii)* capture the likelihood of deep convection in the region. These thermodynamic constraints capture variations in genesis activity on seasonal and intraseasonal (for example, due to the Madden Julian Oscillation) timescales. In contrast, conditions *(iv)-(vi)* track the daily likelihood of genesis (McBride and Zehr 1981); these conditions identify the presence of a weak incipient disturbance (Figure 4.1) and measure how favorable the environment is for intensification: given that the thermodynamics are favorable, genesis is much more likely in an environment with weak vertical wind shear.

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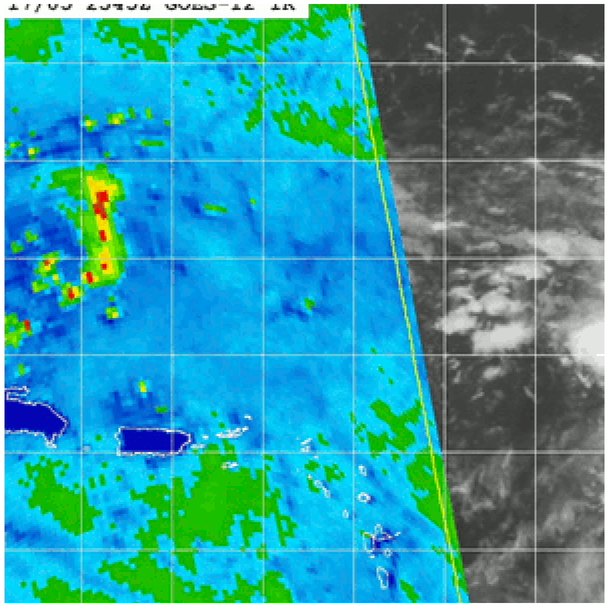

Figure 4.1. Example of an incipient disturbance (to the left of the image) in the North Atlantic: Tropical Invest 96L, which developed into Hurricane Rita (2005). Image is combined infrared and SSM/I satellite images about 0000 UTC 18 Sep 2005. Minimum surface pressure was estimated as 1012 hPa and peak surface winds were around 12 ms-1 at this time.

As observing systems have improved, we have begun to record tropical cyclones that can spend large fractions of their lifetimes within 5 of the equator. These are typically small tropical cyclones (or midget typhoons) and never cross the equator. Often these systems will form from equatorial waves, maybe even a pair of tropical cyclones straddled across the equator. All of this means that restriction on latitude identified by Gray (1968) must be relaxed. We will discuss the different sources for these incipient disturbances in each ocean basin below.

As we've seen then, Gray's (1968) necessary conditions for genesis can be summarized as (1) environmental support for active deep convection [moist free troposphere and weak shear] and (2) in the presence of a low-level absolute vorticity maximum (Evans 2011).

While all of these conditions for genesis must be present before tropical cyclogenesis can occur, even if all of these conditions are met tropical cyclogenesis may not occur. So in that sense, these are necessary, but not sufficient conditions for genesis. The additional factor that guarantees that genesis will occur has yet to be identified, so genesis remains a challenging forecast problem.

4.3 Tropical cyclone structure needed for genesis and intensification to continue

The persistence (and possible intensification) of a weak tropical disturbance depends on its structure. We can use two concepts to understand how a tropical disturbance can survive and grow: inertial stability tells us what a system needs in order to survive in an environment that is not ideal for development; and Rossby radius of deformation tells us if a system has "the right stuff" to sustain itself without help from its environment.

Inertial stability, I

The resistance of a symmetric vortex to changes in its structure can be measured through its inertial stability. Inertial stability indicates the "stiffness" of a cyclone and is calculated as the product of the absolute vorticity and the absolute rotation rate of a symmetric storm, so we see that the *inertial stability varies with radius and height in the storm*[3] (Figure 4.2). This means that the storm can have "weak spots" where adverse environmental influences can have a big impact. In fact, even an intense tropical cyclone has its Achilles' heels in the upper level outflow anticyclone and at the surface (Figure 4.2).

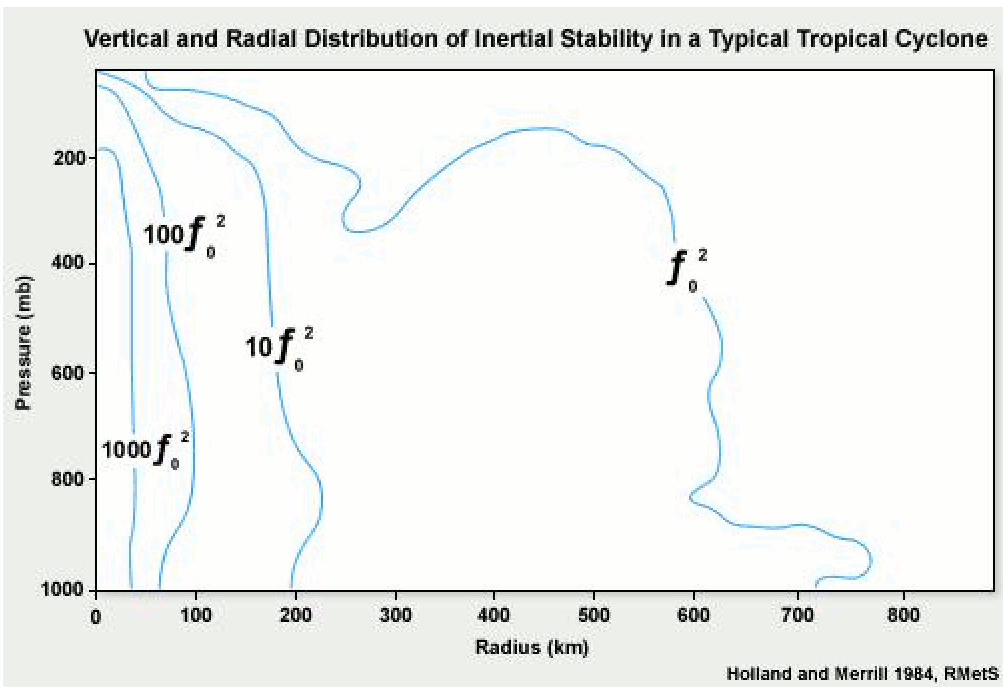

Figure 4.2. Vertical and radial distribution of the inertial stability in a typical tropical cyclone. To illustrate the contribution to inertial stability of the storm winds compared to its environment, the inertial stability values have been scaled by *f0*, the value of the Coriolis parameter at the storm center. Figure adapted by COMET from Holland and Merrill (1984) and obtained from http://www.meted.ucar.edu/tropical/textbook/ch10/tropcyclone_10_2_2.html.

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3Inertial stability is calculated as $I = \sqrt{(\zeta + f_o)\left(f_o + \frac{2v}{r}\right)}$ where ζ is the relative vorticity calculated from the symmetric rotating winds (*v*), *f0* is the Coriolis parameter and *v/r* is the rotation rate of the storm. Therefore, inertial stability increases as relative vorticity, rotation rate and latitude increase, and as radius decreases. As a result, even if a storm has constant intensity, if it is contracting it will increase its inertial stability. This will make the storm more resistant to any negative forcings in its environment.

Rossby radius of deformation, L_R

The Rossby radius of deformation is a length scale that indicates whether convection will force changes in the windfield or if the winds will adapt to the convection — it is the critical length scale at which rotation becomes as important as buoyancy.

Whether a convective system lies above or below the Rossby radius line will determine its future: a disturbance that is larger ($L > L_R$) will persist, but a system that is smaller than L_R will disperse (Figure 4.3). L_R depends on the vorticity, stability and depth of the system[4] and is inversely proportional to latitude; this latitudinal dependence means that it is typically very large in the tropics. However, the high vorticity in tropical cyclones reduces the Rossby radius and enables tropical cyclones to develop and even to last for many days to weeks. Dynamically, the low-level vorticity maximum reduces the local Rossby radius of deformation focusing the convective heating locally (Simpson et al. 1998). This means that the tropical cyclone can intensify more efficiently.

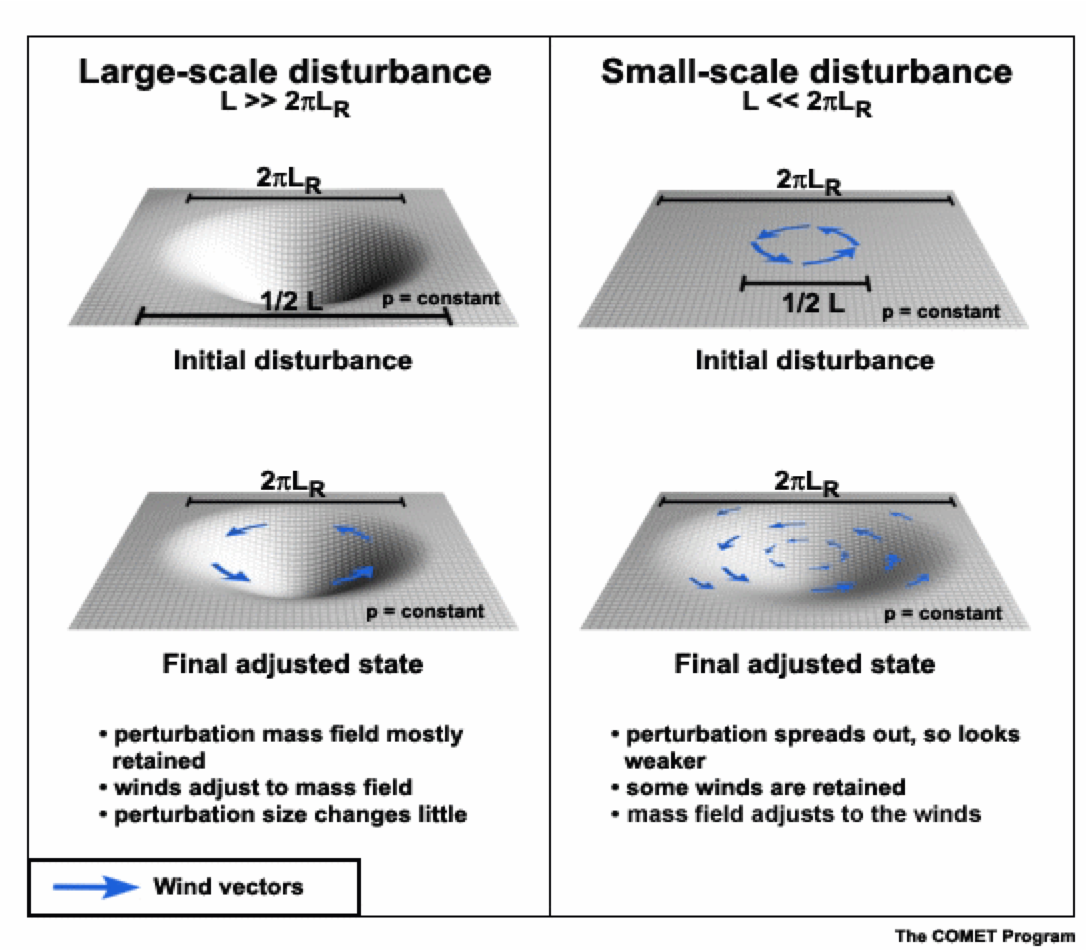

Figure 4.3. Adjustments of the wind and mass (pressure, geopotential height) fields based on the size of the disturbance relative to the Rossby radius of deformation, LR. A small convective disturbance will decay, but a disturbance that is larger than LR can develop and grow.

4The Rossby radius of deformation for a continuously stratified fluid, $L_R = \frac{N\,H}{f_o}$, where *N* is the Brunt Väisäla frequency, *H* is the depth of the system and *f0* is the Coriolis parameter; if a weather system has large relative vorticity (e.g. a tropical cyclone!), the vorticity of the system has to be included and LR is calculated as $L_R = \frac{N\,H}{\zeta + f_o}$ where ζ is the vertical component of the relative vorticity. For more information on the Rossby radius of deformation, see the COMET module, The Balancing Act of Geostrophic Adjustment, http://www.meted.ucar.edu/nwp/pcu1/d_adjust/.

4.4 Regional sources of disturbances that can develop into tropical cyclones

The environments in the eastern North Pacific and North Atlantic are quite different from the other basins where tropical cyclones occur, so some of the sources of the tropical disturbances also differ across basins. In this section we'll review the sources for the tropical disturbances that ultimately become tropical cyclones that are well known (Figure 4.4, bottom panel) and newly identified in the last twenty years or so (top panel).

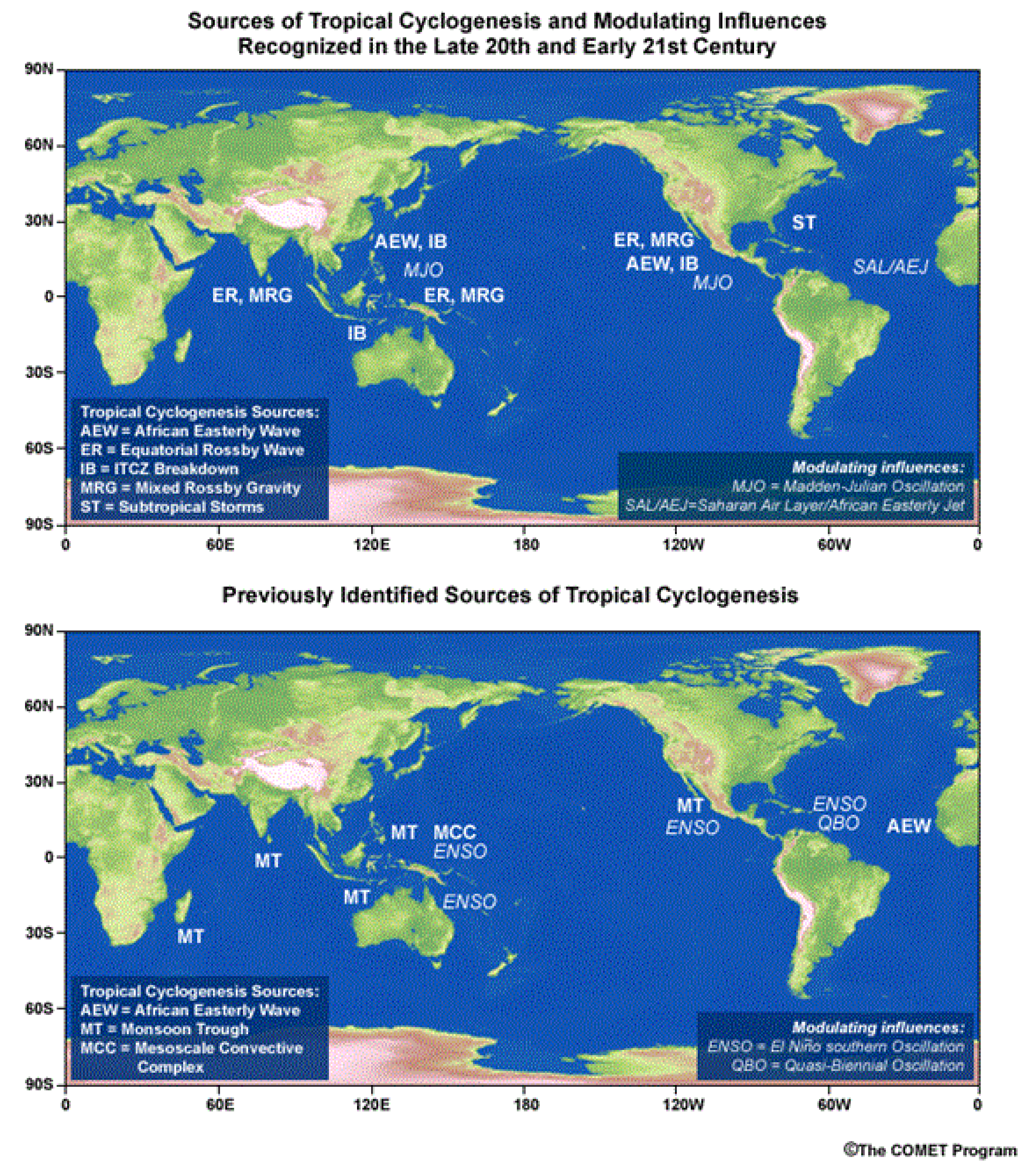

Figure 4.4. Mechanisms for formation of tropical disturbances in each ocean basin. Figures from An *Introduction to Tropical Meteorology* (COMET/UCAR), available from http://www.meted.ucar.edu/tropical/textbook/ch10/tropcyclone_10_3_5.html.

The following mechanisms (shown in Figure 4.4) are currently thought to provide most of the initial tropical disturbances that eventually become tropical cyclones in each basin:

Eastern North Pacific

Tropical storms in the eastern North Pacific mainly form from either instabilities in the ITCZ (Schubert et al. 1991; Ferreira and Schubert 1997) or easterly waves that cross Central America from the Atlantic (Zehnder 1991).

Western Pacific and Indian Oceans

For the West Pacific and Indian Oceans, the summer monsoons are the major source for disturbances that evolve into tropical cyclones (McBride and Keenan 1982; Briegel and Frank 1997). However, recently equatorial Rossby waves (Frank and Roundy 2006) and mixed Rossby gravity waves (Dickinson and Molinari 2002) have also been demonstrated to be other locations for tropical cyclogenesis in these oceans. Finally, some genesis events in the western North Pacific can result from (i) easterly waves from the North Atlantic and (ii) merger of a number of weak mesoscale convective systems into a stronger disturbance (Simpson et al. 1997).

North Atlantic Ocean

Easterly waves forming in the West African monsoon are the major source of Atlantic tropical cyclogenesis. An average of around 1-2 tropical cyclones per year from an initially subtropical cyclone (Guishard et al. 2009).

South Atlantic Ocean

Historically it was thought that the cool SST,lack of a true seasonal monsoon and strong shear in this region meant that tropical cyclones could not form. However, one tropical cyclone has been observed in the South Atlantic, Hurricane Catarina in 2003. Catarina (named after the region in Brazil where it made landfall) evolved from a subtropical cyclone much further from the equator than is typical in other basins.

4.5 Mechanisms for generation of tropical disturbances

We'll now take a very quick look at how each of these mechanisms lead to the formation of the tropical disturbances that eventually becomes tropical cyclones.

The monsoon trough

Convection can strengthen the monsoon trough, progressively lowering the surface pressure and strengthening the winds.At some point, the trough becomes too strong and narrow, and the monsoon can break apart into a set of cyclonic disturbances that may evolve into tropical cyclones (Ferreira and Schubert 1997, Figure 4.5). An isolated monsoon gyre forming in this environment can also provide the seed for a tropical cyclone.

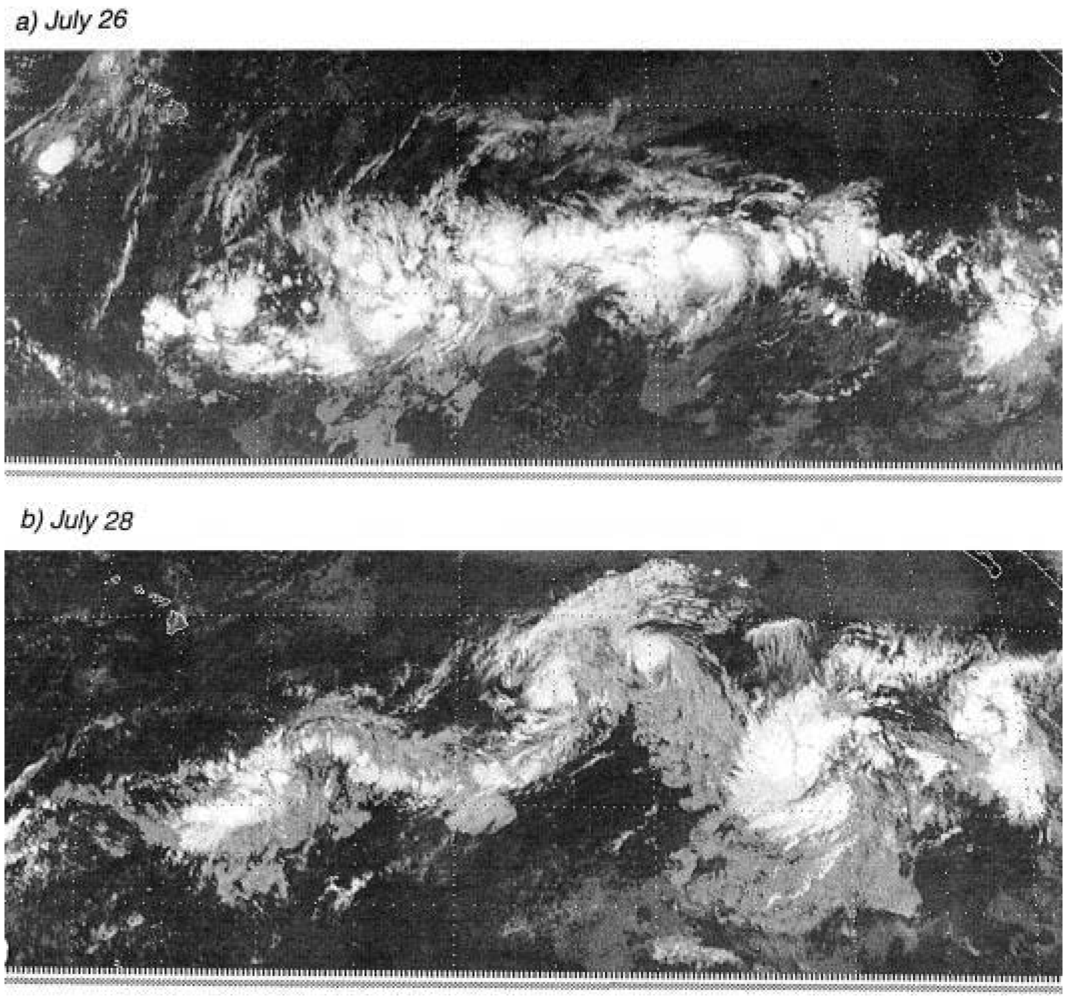

Figure 4.5. Observed breakdown of the continuous monsoon trough (top) into three tropical cyclones (3rd panel) over a period of a week (Ferreira and Schubert 1997) and the beginning of the reformation of the trough structure (bottom panel).

Equatorial waves: Rossby waves and mixed Rossby-gravity waves[5]

Two kinds of atmospheric waves that propagate along the equator have been identified as favorable regions for tropical cyclogenesis: equatorial Rossby waves (Frank and Roundy 2006) and mixed Rossby gravity waves (Dickinson and Molinari 2002). Because of convection, their real structure is different from theory, so it is only recently that we've been able to track these waves reliably and to observe their impact on tropical cyclone formation (Kiladis and Wheeler 1995; Frank and Roundy 2006).

Equatorial Rossby waves are symmetric about the equator, so the cyclonic (clockwise) flow in the wave south of the equator will be mirrored by the (counterclockwise) cyclonic flow to the north of the equator. In an equatorial Rossby wave, these cyclone pairs alternate east-west with pairs of anticyclones. The low pressure centers are regions of active convection, another reason why equatorial Rossby waves are favorable sites for genesis (e.g. Nitta 1989 and also see figure 4.6).

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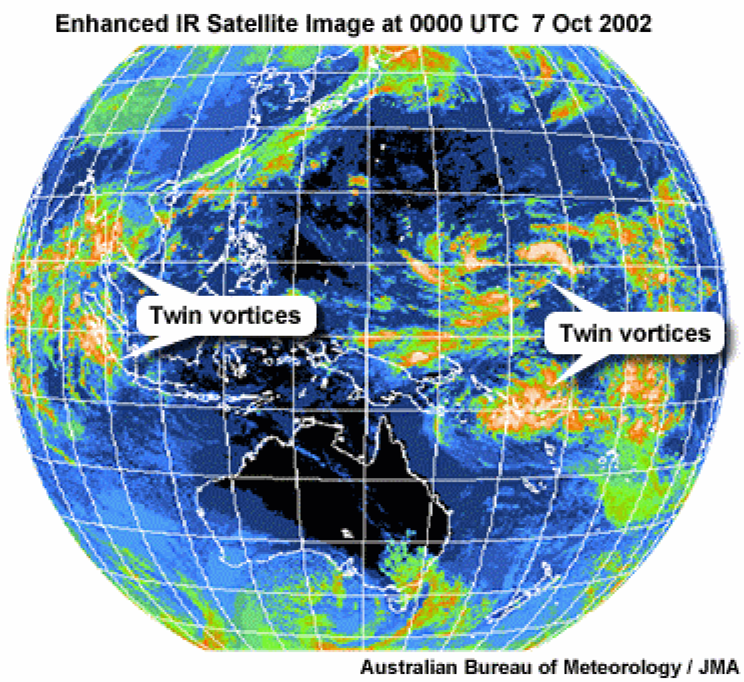

Figure 4.6. An equatorial Rossby wave is characterized by pairs of convective cyclonic regions (favorable for tropical cyclogenesis) alternating around the equator with clear zones of the similar scale. Figure obtained from http://www.meted.ucar.edu/tropical/textbook/ch5/tropvar_5_1_2_2.html.

Mixed Rossby-gravity waves are much shorter than equatorial Rossby waves and only persist for a few days instead of weeks. Even so, they have been shown to be another important source region for tropical cyclogenesis (Dickinson and Molinari 2002, Figure 4.7).

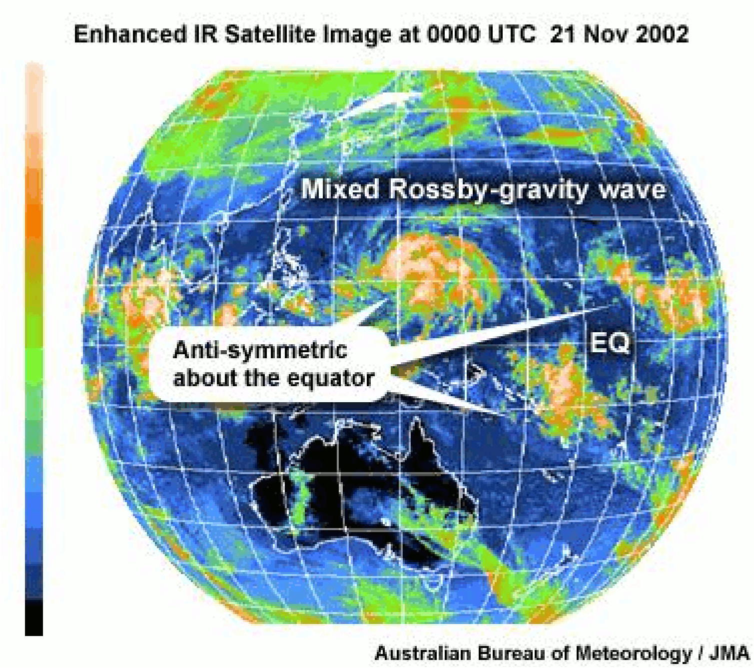

Figure 4.7. An equatorial mixed Rossby-gravity wave observed in the enhanced IR imagery from the GMS satellite at 0000 UTC 21 November 2002. Notice the convective maxima in the cyclonic flow regions and the cyclone-anticyclone pairs straddling the equator (leading to a "sawtooth" pattern of convective cyclonic centers alternating across the equator). Figure obtained from http://www.meted.ucar.edu/tropical/textbook/ch5/tropvar_5_1_2_3.html.

5More information on equatorial waves can be found at Matthew Wheeler's excellent operational websitehttp://www.cawcr.gov.au/staff/mwheeler/maproom/OLR_modes/ or from Chapter 5 (Tropical Variability) the COMET textbook site at http://www.meted.ucar.edu/tropical/textbook/ch5/tropvar_5_1_2.html.

Tropical cyclogenesis from an equatorial wave does not have to occur right near the equator, although it often does. The wave also sees the ITCZ as a "vorticity equator" and will propagate along the ITCZ until it reaches the easternmost point of the monsoon trough. This is another particularly favorable region for tropical cyclogenesis to occur in the western North Pacific since low-level convergence is persistent here, making this an active region of deep convection.

African easterly waves (AEW)

African easterly waves are generated in the monsoon region in western equatorial Africa. They form as instabilities on the African Easterly Jet (AEJ) and will develop if they can form persist convection once they move over the tropical Atlantic Ocean. The development of the AEW from the jet is somewhat similar to the monsoon trough breakdown mechanism we discussed in association with Figure 4.4.

Easterly waves are strongest around 700 hPa (about the height of the AEJ). They have periods of roughly 3-5 days and scales of about 1000 km; AEW typically move westward at about 7-8 m s^{-1}. If an AEW does not develop into a tropical storm close to the African coast, strong vertical wind shear in the central equatorial Atlantic will usually suppress its development until it nears the Caribbean Sea. It may either form a tropical storm here, or just gain energy to continue to move into the eastern Pacific (Molinari et al. 1997)

Mesoscale convective systems (MCS) as sources of tropical disturbances

Mesoscale convective systems (MCS) are organized clusters of convection. Individual MCS are usually too small (compared to the Rossby radius of deformation) to continue to develop into a tropical storm. They are also too weak (low values of inertial stability) to resist the negative influences of their environment. So to become a tropical storm, MCS must band together and merge into a stronger system (larger low-level relative vorticity). At this stage, the Rossby radius (L_R) of the developing storm has reduced so the system can strengthen further through the feedback of convective heating on the rotating winds as depicted in Figure 4.3 (Ritchie and Holland 1999).

Subtropical storms

Although they are listed now and then in the North Atlantic historical database (HURDAT), the role of subtropical storms in tropical cyclogenesis received little attention. Their regular occurrence as a seed disturbance for tropical cyclones has been documented over the last decade (e.g. Davis and Bosart 2003, 2004; Evans and Guishard 2009) and, as a result, the US National Hurricane Center (NHC) began naming those storms and recording them in HURDAT in 2002.

A subtropical cyclone evolves into a tropical cyclone via a process of tropical transition when the shallow, subtropical system (strongest winds in the lower to mid troposphere) evolves into a deep tropical cyclone with strongest winds at the surface). For tropical transition to occur the subtropical cyclone must be in a moist, low-shear environment; it must either be over warm ocean waters (SST > 26°C or more) or in a region of strong warm (and moist) air advection. All of these conditions will force deep convection in this initially baroclinic storm, allowing it to develop the deep warm-cored structure with strongest winds at the surface — a tropical cyclone!

So we see that the development of a subtropical storm into a tropical cyclone requires weak vertical wind shear in a moist, convective environment and in the presence of a relative vorticity maximum. Thus, the necessary but not sufficient conditions for tropical cyclogenesis must still be satisfied in this tropical transition path to tropical cyclogenesis!

4.6 Favorable conditions for the evolution of a tropical disturbance into a tropical storm

No matter how the tropical disturbance develops, it needs a favorable environment (Gray 1968) to continue developing and to intensify. As we've discussed above, this weak tropical system will generally weaken and die in regions of even low to moderate vertical wind shear if it cannot sustain active convection since the shear will tilt the vortex away from the vertical, suppressing vertical motion in the cyclone core – and so suppressing convection that provides the engine for the cyclone. However, sometimes a little shear can be a good thing: in causing the vortex to tilt, it will generate upward motion downstream of the center (under the "upper" vortex) and (if there is enough moisture in the atmosphere) deep convection can become active there. This convection can form a new center and capture the original vortex (Molinari et al. 2004) in the same way as we discussed above for the MCS path to genesis.

Increasing study of variability in the large-scale tropical environment has highlighted the importance of the *Madden-Julian Oscillation* (MJO; Madden and Julian 1994) in modulating deep convection on intraseasonal timescales.

We've repeatedly emphasized the need for active deep convection to feed the engine of tropical cyclogenesis and development, so the impact of the MJO on tropical cyclogenesis is hardly surprising: the active convection phase of the MJO enhances the chances of genesis, while the suppressed phase of the MJO minimizes the likelihood that genesis will occur (Liebmann et al.1994; Molinari and Vollaro 2000; Maloney and Hartmann 2001; Aiyyer and Molinari 2008). This effect is seen in all ocean basins, even the North Atlantic (Evans and Jaskiewicz 2001), where the convective signature of the MJO itself is not readily identified.

One message comes out of our discussion of tropical cyclogenesis: each one of Gray's (1968) necessary conditions for tropical cyclogenesis is present, no matter which path to genesis (monsoon, equatorial wave, subtropical storm, MCS) the storm takes.

4.7 Tropical cyclone intensity and mechanisms of intensity change

Intensity and structure are the key characteristics of a tropical cyclone that determine its impact on society.Together they describe the maximum winds experienced and the distribution of dangerous winds and other significant weather associated with the storm.

To understand and forecast the intensity of a storm,it is helpful to understand the maximum possible intensity that a storm could achieve in its environment: its *potential intensity (PI)*. Having gauged the upper bound on the destructiveness of that storm, a review of the environmental constraints limiting its development provide guidance on the actual intensity it could achieve; the effects of storm structure changes on intensity will also be discussed. Finally, methods for observing, estimating, and classifying storms based on their intensity will be reviewed.

How strong can it possibly get? The potential intensity (PI) of a tropical Cyclone

While the actual intensity attained by a tropical cyclone is a critical measure of its societal impact, the potential intensity of a storm tells us what the "worst case" outcome of that storm could be. Potential intensity is the maximum possible surface wind speed or minimum central pressure attainable by an individual storm given the constraints imposed on it by the thermodynamics of its environment.So potential intensity can help to guide a forecast since it provides a situation-dependent limit on the forecasted intensity (De Maria et al. 2005). Potential intensity is also of interest as an indicator of the possible impacts of climate changes on these storms (Emanuel 1988).

Two theories have been advanced to explain potential intensity,one that links intensity to the characteristics of the moist boundary layer (CISK; Ooyama 1963, 1982; Charney and Eliassen 1964; Fraedrich and McBride 1989) and the other that ties intensity more directly to the underlying ocean (WISHE; Emanuel 1986,1995). In both theories, the latent heat released in the eyewall convection is the energy source needed to maintain the winds. Each theory is summarized here.

Conditional Instability of the Second Kind (CISK), an early theory of potential intensity

(1958) proposed a theoretical model for the maximum intensity of a tropical cyclone dependent on (i) SST at the center of the storm,(ii) relative humidity of the surface air in the storm; (iii)lapse rates in the environment of the storm, and (iv) potential temperature at the top of the storm and its height above the surface. The first three conditions combine to give the moist instability of the air, and so the vigor and depth of the convection in the core. Miller (1958) calculated the hydrostatic pressure drop due to the potential temperature anomaly aloft due to the eyewall convection; his analyses demonstrated that subsidence warming in the eye must also be included to attain the range of observed minimum central pressures.

Conditional instability of the second kind (CISK) is a formal mathematical theory[6] for tropical cyclone maintenance and intensification (Ooyama 1963, 1982; Charney and Eliassen 1964; Fraedrich and McBride 1989) that is consistent with Miller's (1958) conceptual model of a tropical cyclone as a moist, frictionally-driven convective "chimney". In CISK, warm, moist (high θ_e) air converging in the tropical cyclone boundary layer provides all of the moisture available to the eyewall convection. Condensation of rain in the convective eyewall provides the latent heat source that is converted to mechanical energy, driving the winds of the tropical cyclone (Figure 4.8).

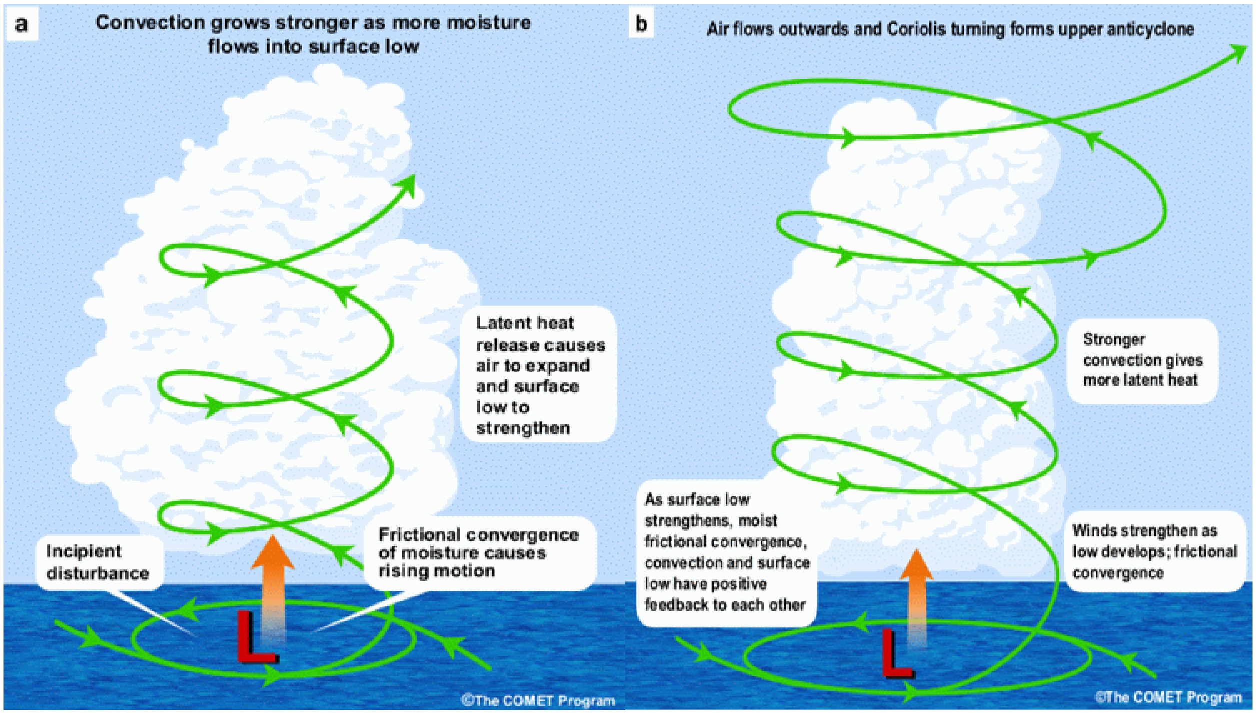

Figure 4.8. Schematic of CISK: (a)given an incipient low-level cyclone with a moist boundary layer, frictional convergence of moisture, and forced ascent drive convection; (b)latent heating due to convection reduces surface pressure, strengthens the low-level cyclone, and enhances moisture convergence and convection in a positive feedback loop. Obtained from http://www.meted.ucar.edu/tropical/textbook_2nd_edition/navmenu.php?tab=9&page=4.1.1.

6Our discussion of these two theories (CISK and WISHE) will be descriptive. Complete mathematical discussions of these theories are available in the source papers; many of these are listed in the references.

Friction in the tropical cyclone boundary layer breaks the gradient wind balance between the inward pressure gradient force and the outward Coriolis and centrifugal forces leads to the boundary layer convergence that imports the moist, boundary layer air needed to maintain the eyewall convection: *so according to CISK, friction is necessary to maintain a tropical cyclone*. Thus, the energy from the latent heat release must at least balance the energy lost to surface friction for a tropical cyclone to maintain its current intensity. If the mechanical energy due to latent heat release *exceeds (is less than)* the energy lost to surface friction, the tropical cyclone will *intensify (weaken)*.

Wind-Induced Surface Heat Exchange (WISHE), a Carnot engine theory of PI

In a Carnot engine, heat energy is converted to mechanical energy. Since tropical cyclones are built on latent heat release from active deep convection, perhaps a Carnot engine could be a good model for a tropical cyclone. The Carnot engine is a closed system in which the temperature difference across the layer provides the thermal energy available to drive motion.[7] A real tropical cyclone is not a closed system,but part of the complicated structure of

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the atmospheric weather. To get around this, Emanuel (1986, 2005) assumed that the air in the upper tropospheric outflow sinking to the surface far from the storm to complete the circuit (Figure 4.9; from C to A). In spite of this and many other simplifying assumptions, the Carnot cycle PI theory provides a reasonable estimate of the upper limit for tropical cyclone intensity in a given environment, although not for the actual intensity at any given time.

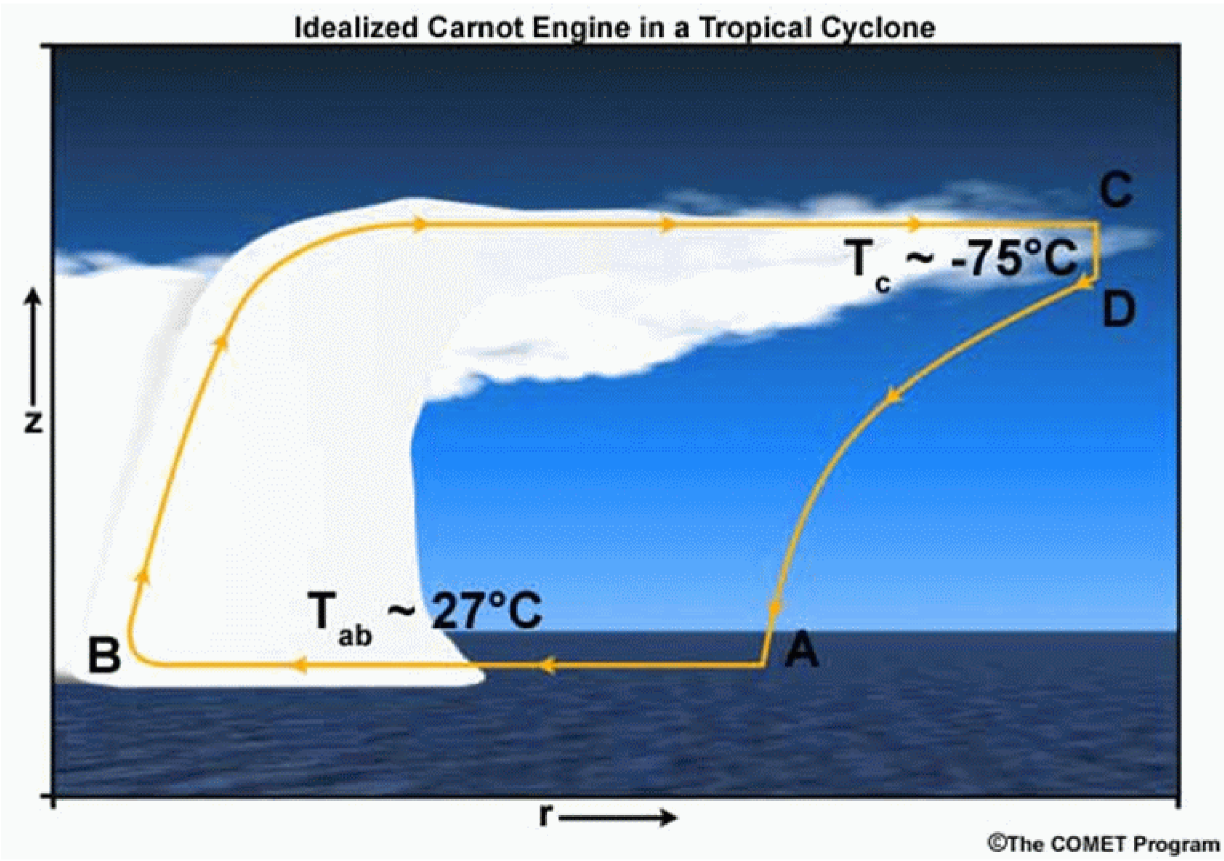

Figure 4.9. Schematic of the heat energy flow in an idealized "Carnot engine" tropical cyclone. Air in the atmospheric boundary layer flows in isothermally (AB), rises adiabatically in the eyewall convection (BC), diverges isothermally near the tropopause in the outflow anticyclone(CD). To close the circuit, this air must sink far from the storm (DA). Obtained from http://www.meted.ucar.edu/tropical/textbook_2nd_edition/navmenu.php?tab=9&page=4.1.2.

7For example, the heating element below a pot of boiling water is much warmer than the air above and the water moves turbulently — in a similar way, the radiative cooling at the top of a cloud drives turbulent motions in the vertical. The strong rotation in a tropical cyclone organizes the motions driven by the latent heat release.

Changes in thermal energy around the Carnot cycle provide the source for the mechanical energy sustaining the winds. Let's follow a parcel around the cycle depicted in Figure 4.9.

Air converging into the tropical cyclone boundary layer (AB) is warmed by (i)sensible heat flux from the ocean surface and (ii)frictional heating due to the deceleration of the winds (Bister and Emanuel 1998); this heating is balanced by (iii) adiabatic expansion as the air flows towards the

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low pressure center and (iv) diabatic cooling due to evaporation(latent heat flux from the ocean). In WISHE, these four processes are assumed to balance, so there is no temperature change for air in the boundary layer inflow branch of the cycle. For calculations in the Carnot cycle model,the temperature of this isothermal inflow is taken to be the SST (Emanuel 2005).

Air rising in the eyewall convection (BC; the upward branch of the Carnot engine) rises along a moist adiabat, so the air at the top of the convective tower has the same saturation equivalent potential temperature[8] as the air feeding these clouds. However, air in the convection becomes increasingly dry as it rains and, in combination with the pressure drop in the vertical, this results in much cooler temperatures in the outflow layer (just below the tropopause) than in the boundary layer inflow. This temperature difference between the ocean surface and the tropopause provides the thermodynamic energy available to be converted into wind energy.

The Carnot engine that approximates our tropical cyclone is not perfectly efficient; only some of the thermal energy is available to be converted into motion. Typical values for a tropical cyclone are SST~27°C and tropopause temperatures about -75°C. Using these we calculate the efficiency, ε, of the Carnot engine[9] to be around one third, meaning that one third of the thermodynamic energy in our Carnot engine tropical cyclone is available to drive the winds.

8Equivalent potential temperature, *θe*, is defined as $\theta_e = T_e \left(\frac{p_0}{p}\right)^{\frac{R_d}{c_p}}$ where $T_e = \left(T + \frac{L_v r}{c_p}\right)\left(\frac{p_0}{p}\right)^{\frac{R_d}{c_p}}$ is the equivalent temperature corresponding to (*T, p, r*) at that location; *T* (K) and *p* (hPa) are the temperature and pressure of the parcel, *r* is the water vapor mixing ratio of the parcel (in units of kg kg-1) and *Lv*, the latent heat of evaporation (J kg), which varies with temperature. The constants appearing in this equation are: *p0*, the reference pressure (1000 hPa); *Rd*, the gas constant for dry air (287 J kg-1 K-1); and *cp*, the specific heat at constant pressure (1004 J kg-1 K-1)

9 $$\varepsilon = \left(\frac{T_{IN} - T_{OUT}}{T_{IN}}\right) \cong \left(\frac{SST - T_{TROP}}{SST}\right) = \left(\frac{300 - 200}{300}\right) = \frac{1}{3}$$

The climatological values of potential intensity derived from this theory agree quite well with the distribution of the most intense tropical cyclones observed (Figure 4.10), so we have some confidence that it can provide guidance on the upper limit for the intensity of an individual tropical cyclone (e.g. De Maria et al. 2005).

However,<u>this WISHE theory is not predictive: it does not tell us what the actual intensity will be for any storm</u>. The evolution of the intensity of an individual tropical cyclone is governed by its complete environment, not just the thermodynamics. Next, we will look at all of the other factors that go into determining the intensities achieved by a real tropical cyclone.

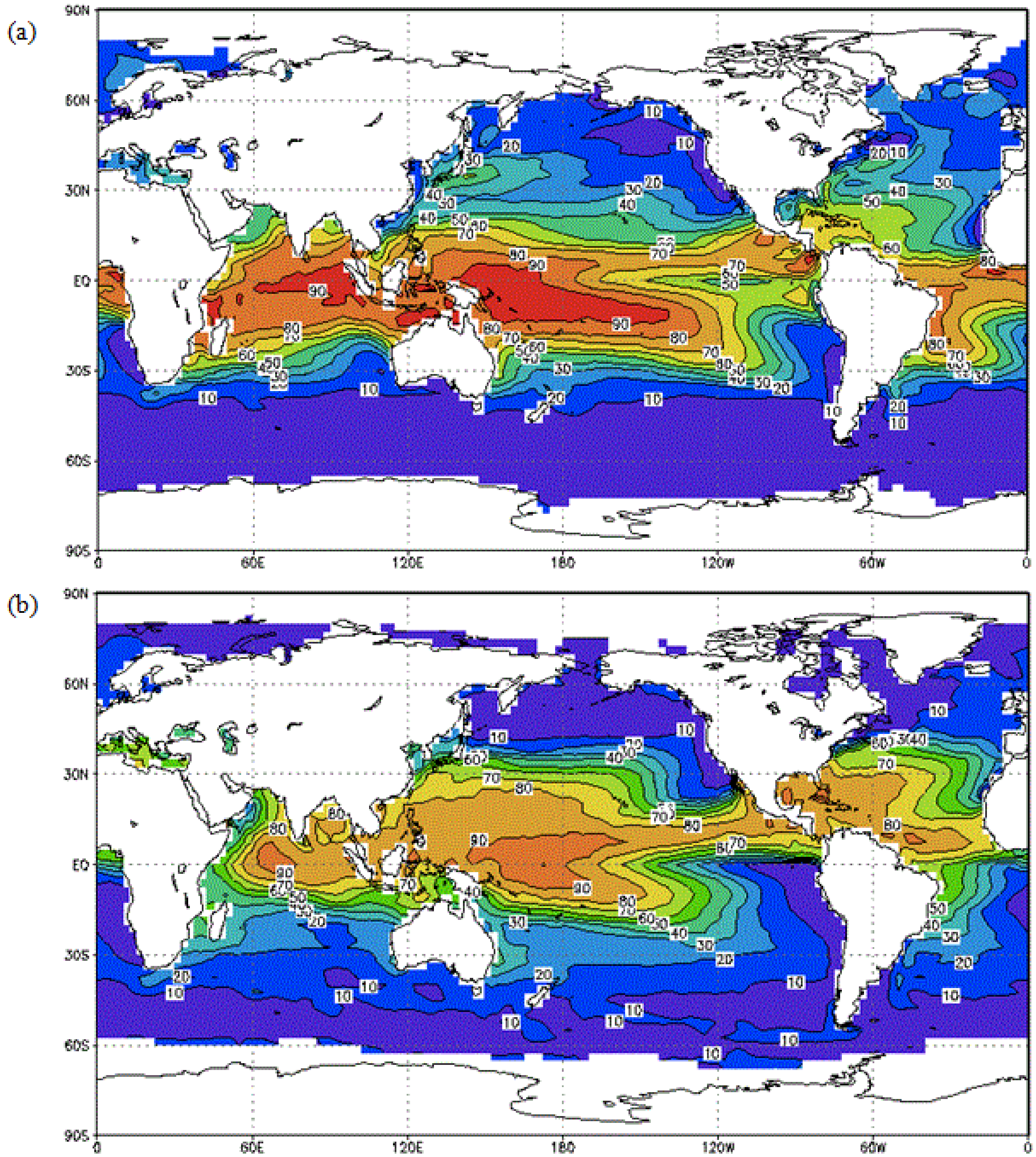

Figure 4.10. Monthly mean maximum surface winds (ms-1) calculated from the WISHE Carnot cycle model using 1982-1995 daily SST and tropopause temperatures: (top) February and (bottom) September. Note the change in color scale between the panels. Maps obtained from http://wind.mit.edu/~emanuel/pcmin/climo.html.

Observed relationships between tropical cyclone intensity and ocean temperatures

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CISK and WISHE theories assume that warm ocean waters,much cooler outflow temperatures, boundary layer convergence of very moist tropical air in an environment supportive of deep convection are readily available to a developing tropical storm. If this were true for all tropical cyclones, tragedies like the Bhola (1970) and Chittagong (1990) tropical cyclones; Typhoon Vera (1959, known as Isewan Typhoon in Japan), Tropical Cyclone Tracy (1974), Hurricanes Mitch (1989) and Katrina (2005) would be the norm. Since these are rare events, it is clear that the large-scale environment of the developing tropical cyclone must interfere with the processes captured in these two theories. What does this mean? It means that the environment must somehow disrupt the deep convection in the evolving tropical cyclone, particularly near its core. Environmental conditions limiting the convection,and so limiting the intensity,of a storm include insufficient moisture (either boundary layer air or relatively dry free troposphere), cool SST,or a region of strong baroclinicity (shear and temperature gradients). While each of these environmental characteristics can inhibit intensification, taking these factors to the other extreme, for example, no vertical wind shear and very warm waters, does not guarantee that a storm will reach its PI. We'll return to this conundrum below.

"Potential" intensity is only achievable when these factors combine favorably. This is demonstrated in a study that compares the observed distribution between intensity and SST (Figure 4.11, reproduced from Evans 1993). A 20-year period chosen (1967-1986) began with the satellite era and ended with the end of aircraft reconnaissance in the western North Pacific.The complete recorded lifecycle for every tropical storm observed was included in the analyses, so these results should be interpreted in terms of the storm lifecycle. The first impression from these analyses is that the two basins with reconnaissance (panels (a) and (b)) show a dramatic increase in intensities for "tropical" SST > 25°C or so, while the strongest storms are restricted to a smaller, warmer range of SST as expected from both the CISK and WISHE theories discussed above. However, we also see that tropical cyclones can achieve any intensity for storms over SST > 25°C.

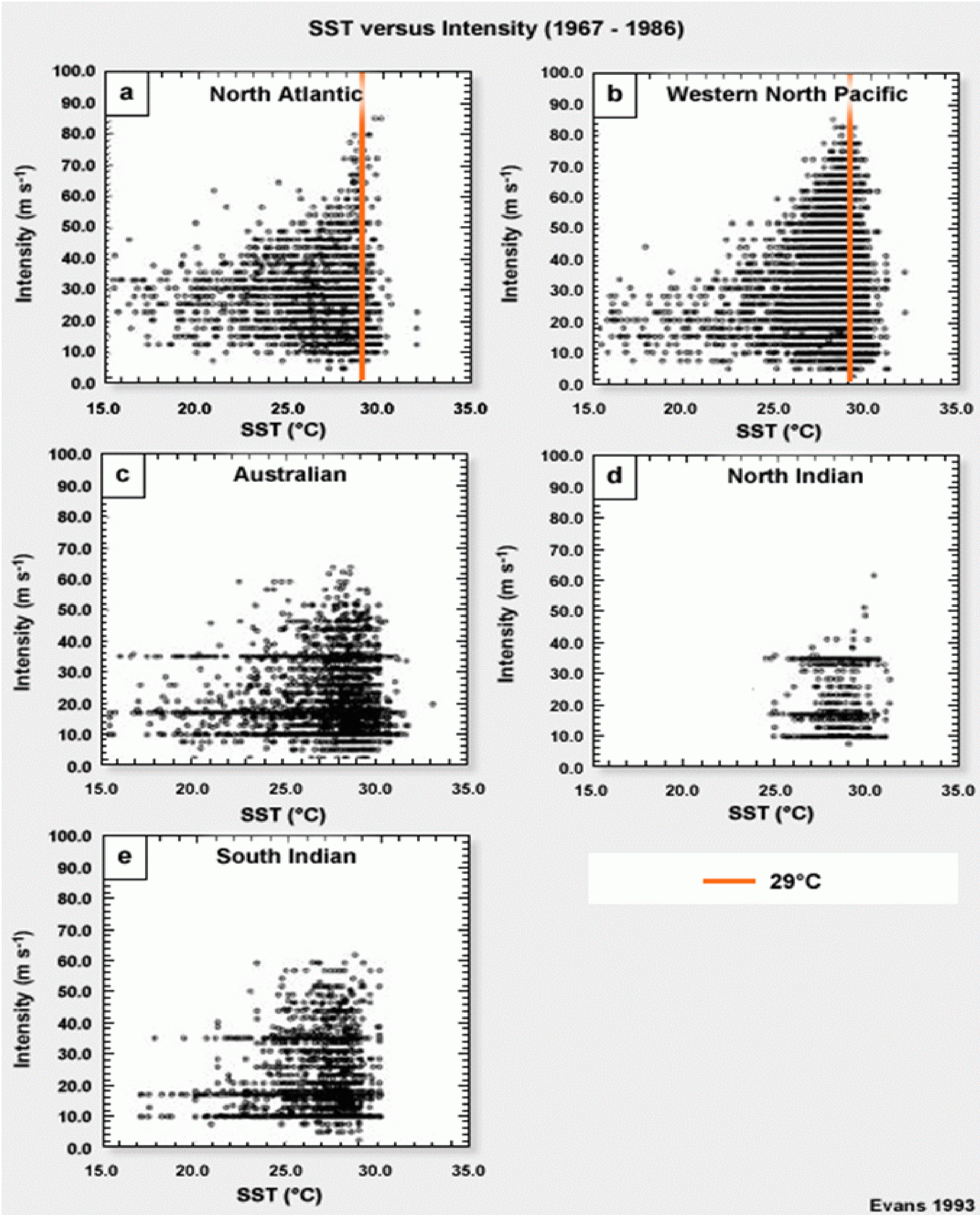

Figure 4.11. Intensity (ms-1) against SST (°C) for all tropical storm times in the 20 year period 1967-1986 for (a) North Atlantic, (b) western North Pacific,(c) Australian,(d)North Indian, and (e) South Indian Ocean basins (Evans 1993, her Fig. 1).

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So what about the tropical ocean basins without regular reconnaissance flights (panels (c)-(e))? We see that the intensity distributions in these regions emphasize key intensity threshold values (likely reflecting a dependence on satellite-based algorithms to estimate intensity). Evans (1993) went on to analyze these data further and demonstrated the expected decrease in intensity with time after genesis, as storms move away from their tropical genesis environment and into cooler waters. When she compared only the peak intensity of each storm against SST, no pattern was evident. These results combined provide strong evidence that warm SST are not sufficient on their own to explain the intensity of a tropical cyclone-reinforcing the role of these two theories in identifying the peak possible intensity, not the actual intensity of a storm.

One more word on SST before we move on. The warm near-equatorial SST is unfavorable for tropical cyclone development since this is also a region of vanishing Coriolis parameter and a region where cyclonic potential vorticity changes sign. As a result, it is difficult to create a balanced, cyclonic incipient disturbance that will intensify into a tropical cyclone. As satellites have improved, storms forming very near the equator (within 5° or so), but never on the equator, have been observed. These are usually very small ("midget") cyclones while they are close to the equator (Harr et al. 1996). Hence, the Earth's vorticity (Coriolis) governs intensity and structure, not just genesis.

Based on these results, we conclude that the organization of the wind field must also be important in determining tropical cyclone intensity. We'll first look at the effects of the environmental winds on intensity, then return to the role of storm structure on its intensity.

Environmental factors limiting tropical cyclone intensity

Interaction of a tropical cyclone with a region of strong environmental vertical wind shear usually leads the storm to begin to weaken, since the storm structure must change as it adjusts to this new environment (Jones 1995, 2000a). Interaction with strong environmental wind shear will initially make the convection of the storm asymmetric (discussed further below). As a result of strong shear,the storm will either (i) reintensify as a tropical system, having generated sufficient convection to retain its tropical structure in the sheared environment; or (ii) begin to recurve, when it will initially weaken (Riehl 1972; Evans and McKinley 1998). Once it begins recurvature, this tropical cyclone will either decay or undergo extratropical transition (discussed in the structure section).

Strong vertical wind shear does not have the same impact on tropical cyclones of different intensities. As we discussed above, as the intensity of a tropical cyclone increases, its inertial stability increases. Since inertial stability is a measure of the resistance of a tropical cyclone to external influences, more intense tropical cyclones are more resilient in adverse environments. For example, a storm rotating with peak surface windspeeds of 50 ms^{-1} would be less disrupted by strong vertical wind shear than a storm of 20 ms^{-1}. A larger tropical cyclone (with strong winds far from its center) will also survive strong shear more easily since it has large values of inertial stability out to larger radius (Merrill 1984).

Two places where even the strongest tropical cyclone is susceptible to its environment are the ocean surface and the outflow anticyclone. Inertial stability in the upper-tropospheric anticyclone is weak, so the tropical cyclone is susceptible to weakening caused by upper-tropospheric weather systems (Hanley et al. 2001). Further, no matter how intense a tropical cyclone is, inertial stability will not protect it from the negative influence of cool SST. Cool ocean waters cut off the deep convection needed for a tropical cyclone to sustain itself (e.g., Figure 4.11).

Now that we've established that strong vertical wind shear limits the intensity of a tropical cyclone, can we conclude that the best environment for a tropical cyclone to intensify is no vertical wind shear? We cannot. Some vertical wind shear is needed to advect the storm along (Frank and Ritchie 1999, 2001). Without this advection, a storm will move very slowly (Chan and Williams 1987); a slow-moving storm will mix cooler ocean water from below the thermocline, cooling the SST below. (Storm rainfall may also contribute to these cooler SST, but this is a small additional effect.) By cooling the underlying SST, the storm weakens the convection in its core, so limiting its own peak intensity. While the shear associated with storm advection may delay intensification, if all other factors are favorable the storm may still reach its PI (Kimball and Evans 2002).

Impacts of tropical cyclone structure on intensity change

As we have discussed,the key to a storm maintaining its current intensity or intensifying further is the maintenance of the deep convection surrounding its core.

As mentioned above, vertical wind shear initially causes the convection in a tropical cyclone to become asymmetric (e.g., Corbosiero and Molinari 2002; Jones 2000b; Molinari et al. 2006). If the convection becomes asymmetric,this lowers the inertial stability of the storm, making it more susceptible to negative effects from its environment.If the shear persists, it can eventually lead to the decay of the tropical storm, as evident with the majority of tropical cyclones that move into the midlatitude baroclinic zone.

As we've said repeatedly, a consistently moist boundary layer is required to feed the tropical cyclone convection; even the convective downdrafts have to be saturated for the convection to be as vigorous as possible in its environment. Evaporation from the ocean must act too for the boundary layer to recover from injection of subsaturated and cool downdraft air; this will take a number of hours before intensification can resume. Intrusions of dry air advected from nearby land masses will also weaken the storm convection. In the North Atlantic,the Saharan Air Layer (SAL) (Carlson and Prospero 1972) also weakens the tropical storm convection (Karyampudi and Pierce 2002; Dunion and Velden 2004) and constrains its motion.

Rapid intensity change: Warm ocean eddies and eyewall replacement cycles

Rapid intensification or weakening are critical challenges when forecasting tropical cyclone intensity evolution. These situations arise predominantly from cyclone passage over a warm ocean eddy or eyewall replacement cycles.

Movement of a tropical cyclone over a warm ocean eddy invigorates the deep convection of the system and so provides extra latent heat energy that is available to intensify the storm.

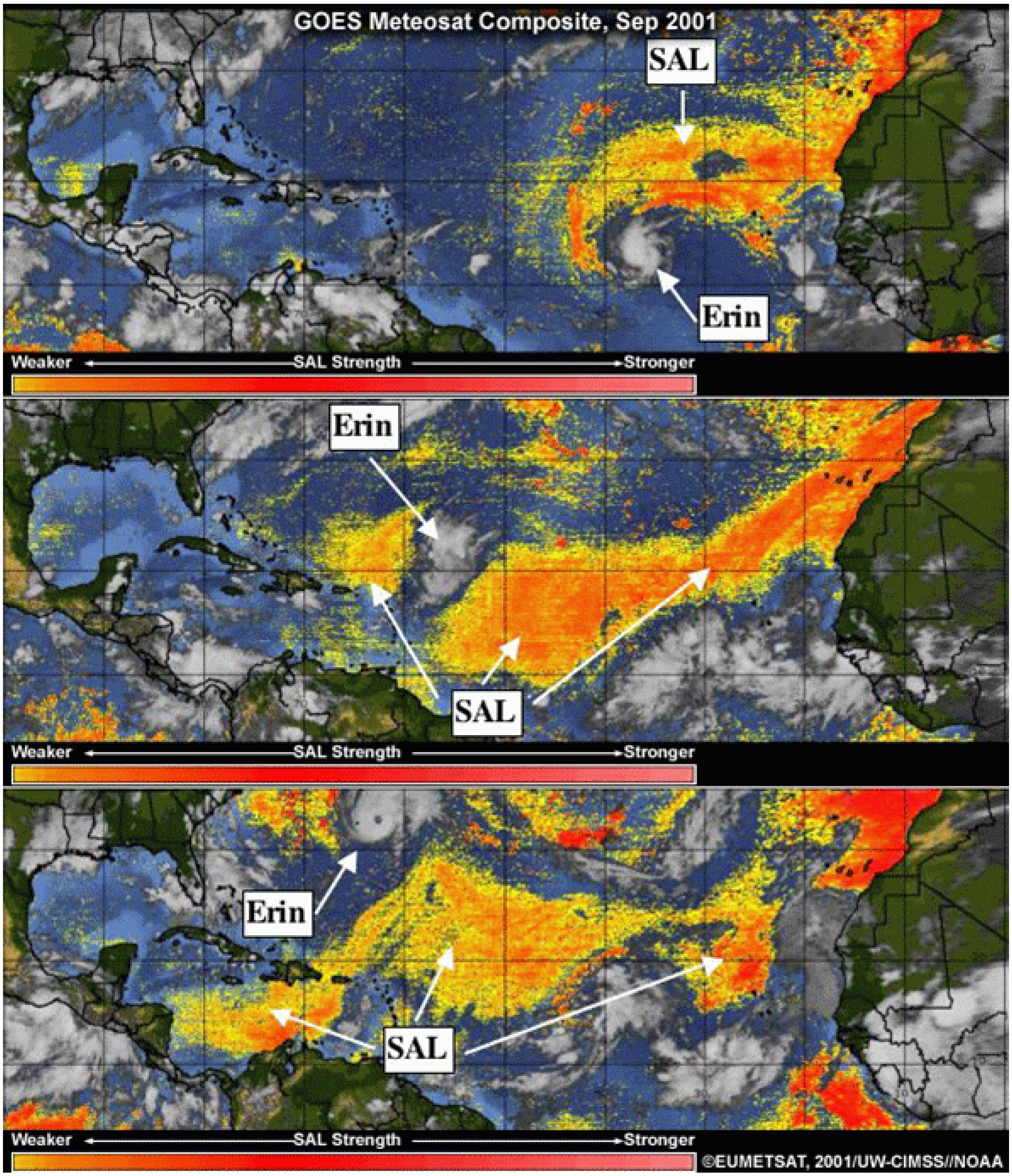

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Figure 4.12. GOES SAL-tracking imagery for the interaction of Hurricane Erin (2001) with a SAL outbreak. Time goes down: (a) 0000 UTC 2 Sep, (b) 1800 UTC 5 Sep, (c) 1800 UTC 9 Sep. Image from http://www.meted.ucar.edu/tropical/textbook/ch10/tropcyclone_10_3_4.html. Real-time diagnostics of the SAL such as these are available from http://tropic.ssec.wisc.edu/

Eyewall replacement cycles occur when an outer eyewall develops, choking off the moist boundary layer inflow needed for the original eyewall to maintain itself; the existing eyewall dissipates and, as a consequence, the central pressure rises (so the surface pressure field of storm relaxes) and the tropical cyclone weakens (Willoughby et al. 1982; Willoughby 1990). At this stage, the outer eyewall contracts gradually and the tropical cyclone regains its original intensity or even becomes more intense (Figure 4.13). Eyewall replacement cycles are most commonly observed during periods of rapid intensification or weakening of intense tropical cyclones (peak winds exceeding 50 ms^{-1}).

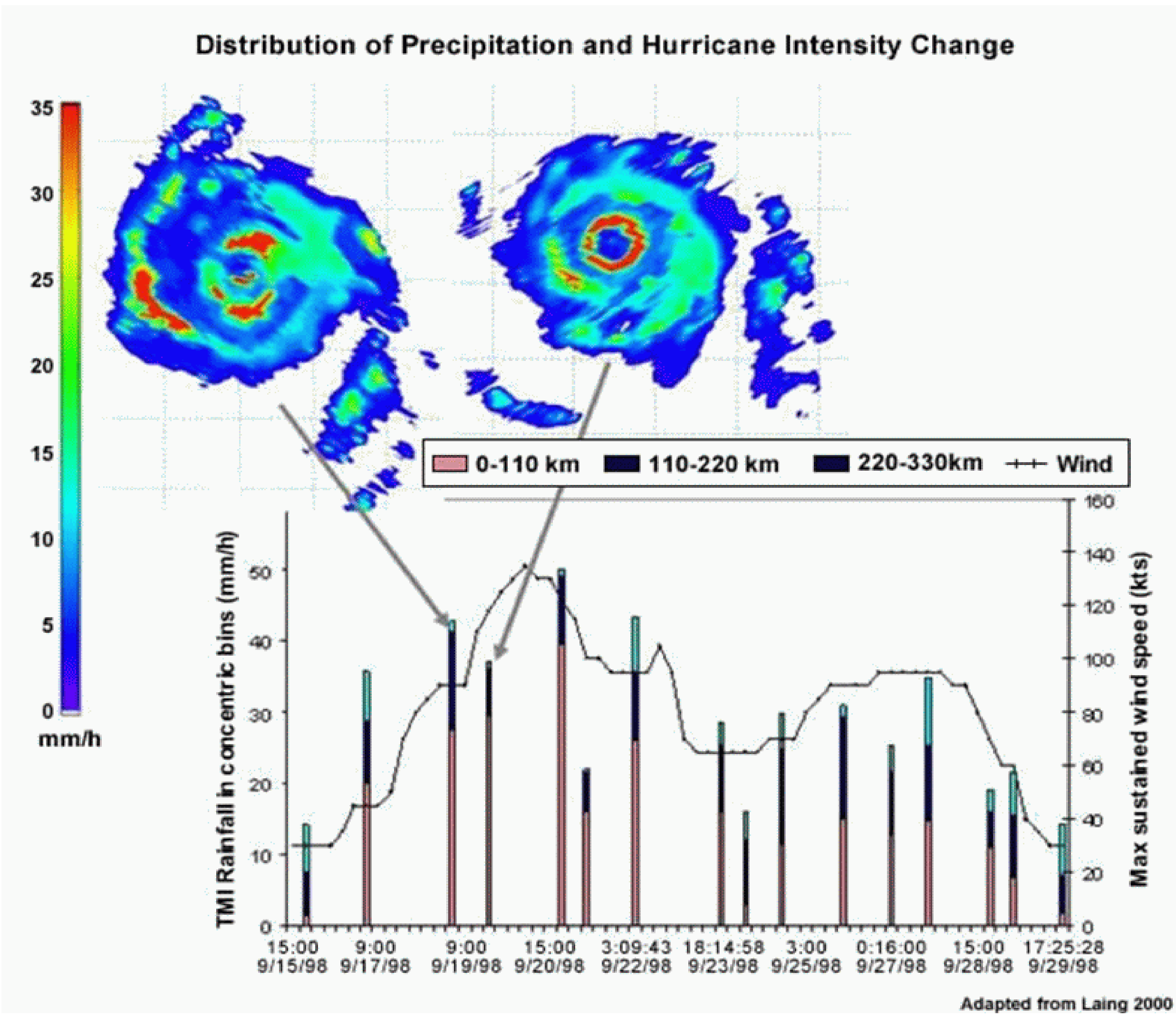

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Figure 4.13. Spatial distribution of surface rain rates (mm hr-1)in Hurricane Georges via TRMM TMI, best track maximum sustained wind speed (knots, solid black line), and rainfall in concentric bins around the center of the cyclone (mm h-1, vertical bars). Laing and Evans (2012).

Concentric eyewalls and eyewall replacements are most common in the western North Pacific because this largest of the ocean basins has very warm waters, meaning that storms typically have longer to intensify before they encounter adverse environmental conditions (such as strong shear,land or cool SST). Some intense tropical cyclones can undergo multiple eyewall replacement cycles.

In summary, intensity change is affected by the current structure and intensity of the tropical cyclone, as well as its environment. Processes leading to intensity change feed back on each other.

Operational monitoring of tropical cyclone intensity

While it is a standard metric of a tropical cyclone and provided in all storm advisories or reports, tropical cyclone intensity is most often deduced via remote sensing[10]. Instruments used for inferring tropical cyclone intensity and structure characteristics include satellite radiometers (IR, visible, and microwave), scatterometers and radars, as well as ground-based radars. Outside special field experiment opportunities, airborne remote sensing of tropical cyclones is only routinely available in the North Atlantic and the eastern North Pacific; in these basins NOAA Hurricane Hunter[11] aircraft are equipped with Doppler radar and stepped frequency microwave radiometers (SFMR). Each of these observing platforms samples different volumes of the atmosphere at different resolutions and accuracy. Thus, different instruments are better suited to measuring different aspects of the storm wind field.

In terms of societal impacts,the peak wind at any single location gives the "worst possible" feature of the storm, so a single wind estimate or small region of intense winds are often be used to determine the observed tropical cyclone intensity (Avila 2010, personal communication).

10A comprehensive discussion of remote sensing applications to tropical cyclone structure and intensity is available from http://www.meted.ucar.edu/tropical/textbook/ch10/tropcyclone_10_4_5.html.

11http://www.aoc.noaa.gov/

Assumptions about the temporal and spatial scales relevant for characterization of intensity are implicit in the calculation of intensity. *The WMO standard definition of intensity is the maximum 10-minute sustained wind at 10 m above the surface.* However, while 10 m above the ground is the standard reference height used for intensity estimation, most operational forecast centers — including WMO regional forecast centers (RFC) — use different time averages. Critical wind speed thresholds — for example, from typhoon to supertyphoon — also differ across basins, as do the names assigned to the storms[12]. These various averaging periods, local classifications and wind speed thresholds are summarized in Tables 4.1 through 4.3 below.

Table 4.1. Intensity classifications and wind speed thresholds for the western North Pacific and Indian Ocean basins.

Western North Pacific	North Indian	South Indian	$m\ s^{-1}$	$km\ h^{-1}$
Tropical Storm	Tropical Storm	Tropical Storm	17-33	60-119
Typhoon	Severe Cyclonic Storm	Tropical Cyclone	> 34	> 120

Table 4.2. Intensity categories and windspeed thresholds for the North Atlantic and Eastern North Pacific Oceans and the Australian Region. See Table 4.3 for the averaging times used for estimating intensity in each basin.

Storm Category	North Atlantic and Eastern North Pacific			Australian Region	
	$m\ s^{-1}$	$km\ h^{-1}$	mph	$m\ s^{-1}$	$km\ h^{-1}$
1	33-42	119-153	74-95	< 34	< 125
2	43-49	154-177	96-110	34-47	125-165
3	50-58	178-209	111-130	47-63	165-225
4	59-69	210-249	131-155	63-78	225-280
5	70+	250+	156+	>78	>280

12 See http://www.wmo.int/pages/prog/www/tcp/Storm-naming.html for background on naming conventions in each basin and lists of upcoming tropical cyclone names by region.

Table 4.3. Intensity averaging times, key wind speed thresholds and naming conventions for all ocean basins impacted by tropical cyclones.

Region	Wind Averaging Time[13]	Special Intensity Threshold		Naming Convention
		$m\ s^{-1}$	$km\ h^{-1}$	
Australia	10 minutes	>35	>125	Tropical Cyclone
		>46	>165	Severe Tropical Cyclone
North Atlantic and Eastern North Pacific	1 minute	>33	>120	Hurricane
		>50	>180	Major Hurricane
Western North Pacific	10 minutes	>63	>227	Super Typhoon
North Indian Ocean	10 minutes	>33	>120	Severe Cyclonic Storm
South Indian	10 minutes	>63	>227	Severe Tropical Cyclone

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[13]A new WMO report http://www.wmo.int/pages/prog/www/tcp/documents/WMO_TD_1555_en.pdf discusses the implications of these various averaging times.

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