Tropical Meteorology

A First Course

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Preface

As the title suggests, this course packet is geared towards students who have had little or no experience to tropical meteorology. However, it is assumed that students have knowledge of basic Calculus, physics, and have taken at least a couple of fundamental meteorology courses. It is strongly advised that students have taken a course in atmospheric thermodynamics, or at least are taking it concurrently.

As this is the second edition of this packet and many changes were made from the first edition, there will inevitably be several errors. Please be a good citizen and let me know where they are so that I may correct them for future generations. Also, I welcome your feedback in terms of course and packet content. If there are topics that you would like to see, or if you feel that there are topics that would be better left out, please let me know.

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"The tempest arose and wearied me so that I knew not where to turn... eyes never beheld seas so high, angry and covered by foam. The wind not only prevented our progress, but offered no opportunity to run behind any headland for shelter...Never did the sky look more terrible; for one whole day and night it blazed like a furnace, and the lightning broke forth with such violence that each time I wondered if it had carried off my spars and sails; the flashes came with such fury and frightfulness that although the ships would be blasted. All this time the water never ceased to fall from the sky; I don't say it rained, because it was like another deluge."

- Christopher Columbus, 1502

“Considering the dire circumstances that we have in New Orleans, virtually a city that has been destroyed, things are going relatively well.”

- Former FEMA Director Michael Brown
  September 1, 2005
Chapter 1
Introduction

The 4 pm Central Daylight Time public advisory issued by the National Hurricane Center (NHC) on August 28 warned of the impending disaster:

“…POTENTIALLY CATASTROPHIC HURRICANE KATRINA HEADED FOR THE NORTHERN GULF COAST…”

It had already been an extremely busy season at NHC, and now there was a tempest swirling in the Gulf of Mexico with 170 mph winds headed straight for every forecaster’s worst nightmare – New Orleans. NHC became a beehive of activity. Media crammed for space and airtime in the media room; forecasters hurried back and forth between computer screens filled with ominous images of Katrina; incoming calls from radio, TV, emergency management officials, and women and men on the street would not cease. But even at that point in time, even with the knowledge of the vulnerability of the city, no one could have anticipated what was to follow over the next few days, weeks, months, and years.

The Hurricane Katrina disaster was a perfect storm of the fury of Mother Nature, bureaucratic ineptitude at many levels, and a city built below sea level. From a meteorological perspective, Katrina was handled very well. The NHC forecast track took Katrina into the New Orleans area 2.5 days before landfall. NHC director Max Mayfield personally phoned New Orleans mayor Ray Nagin to warn him of the situation – something that Mayfield had never done before.

Once the societal impacts of Katrina are scraped away, one begins to wonder about the nature of the tropical cyclone itself. How does an area of the world, known for its gentle trade winds and warm atmosphere, produce a tropical cyclone that has enough kinetic energy (~ 1.5 x 10^{12} Watts) to account for about one half of the world’s energy creating capacity?

There are other mysteries hidden in the tropics that we will explore in this class. What causes the El Niño-Southern Oscillation (ENSO)? How does ENSO manage to impact weather patterns around the world? How do tropical oscillations such as the Madden-Julian oscillation (MJO) provide “triggers” for both ENSO and tropical cyclone formation? Why do tropical cyclones form in some areas of the tropics but not others? How do the tropics force the global circulation?

And the questions don’t end. Tropical meteorology is a fascinating topic for anyone interested in meteorological phenomena at a variety of scales: microscopic (air-sea interaction), mesoscale (convective storms), synoptic (tropical cyclones), and planetary (general circulation). The tropics are unique, in that the Coriolis force is very small there. This presents a whole new set of behaviors that differ from the mid-latitude dynamics that we are used to. Atmospheric temperature gradients are very weak, thus tropical phenomena are much more influenced by things like latent heat release than baroclinicity. The tropics, if defined as the area between 30°N and 30°S, encompass about ½ of the earth-atmosphere system. The size of the tropics, in conjunction with the vast energy surplus accumulated there, has a profound influence on mid-latitude weather.
There are several other important differences between the tropics and the extra-tropics (areas outside of the tropics):

1. Tropics have higher temperature and moisture content
2. Tropics are more unstable – lower temperatures in the upper troposphere
3. Weather systems generally move from east to west in the tropics
4. Precipitation forms primarily through the collision-coalescence process
5. Data availability is sparse
6. Tropics are less predictable, since there is no dominant wave motion like synoptic-scale storms in the mid-latitudes.

Several sections of these notes will deal with those items. In particular, we will address items nos. 1, 2, 3, and 5 in more detail. However, the first few sections will focus on the topic of tropical cyclones, obviously the most visible and violent product of the tropics.
Section I: Tropical Cyclones

Chapter 2
Tropical Cyclone Climatology

Tropical cyclones form in all tropical ocean basins with the exception of the Southeastern Pacific. Gray (1968) published a classic paper on global tropical cyclone formation. Although the paper is several decades old, most of the information is still relevant today. That paper will provide the foundation for this section.

Figure 1 shows the locations of known tropical cyclone formations during an extended time period in the early to mid 20th century (there is no detectable change in genesis locations since that time). The primary genesis areas are: northern and southern Indian Ocean, northern and southern western Pacific, northeastern Pacific, and northern Atlantic. There are no genesis points shown in the southern Atlantic, but there has been tropical cyclone genesis in that area quite recently (Hurricane “Catarina” slammed into Brazil in 2004, killing 3 and injuring 38). TC formation is limited by cool SSTs in the south Atlantic and Pacific.

![Figure 1. Location points of first detection of disturbances which later became tropical storms (From Gray 1968, Copyright American Meteorological Society).](image)

Favored regions in Figure 1 share some common characteristics: warm SST, favorable vertical wind shear, significant Coriolis force, and a generally moist troposphere. Note that there is no genesis activity near the Equator despite the warm SSTs usually present there.

The NW Pacific (36%) has the highest percentage globally of TC formation (Figure 1). This region is followed by the NE Pacific (~16%), the Atlantic and S. Pacific (11% each), and the Bay of Bengal and Indian Ocean (10% each). These numbers may be somewhat inaccurate, since many storms in the pre-satellite era of the early part of the 20th century (especially in areas like the NE Pacific) were probably missed.

Besides sufficient planetary vorticity, the primary factor that limits TC formation over warm water is excess vertical wind shear. Usually, vertical wind shear is measured as the vector difference in winds between the 850 mb and 200 mb levels. For TC genesis to be successful, latent heat must be able to accumulate in the column over the developing storm center. This allows for further pressure drops at the surface, leading to enhanced low-level inflow and increasing latent heat release. High wind shear disrupts this process by displacing the latent heat
away from the surface circulation. Intensification is stopped and, if the wind shear is great enough, the system may be blown apart in the vertical.

Figure 2. Global TC development percentage by basin. 26.5° isotherm is shown (From Gray 1968, Copyright American Meteorological Society).

Figure 3 shows the climatological vertical wind shear for January (which is a prime development month for the Southern Hemisphere) and August (Northern Hemisphere development month). In January, wind shear over the northern Atlantic basin is 40-60 kt, much too high for any development to occur (SSTs are warm enough in the tropical Atlantic, even in January). The more favorable regions of the world are in the western Pacific and Indian Oceans. During the month of August, vertical wind shear decreases.

Figure 3. Climatological average for January (top) and August (bottom) of the zonal vertical wind shear between 200 mb and 850 mb. Positive values indicate the zonal wind at 200 mb is stronger from the west or weaker from the east than zonal wind at 850 mb. Units are in knots (From Gray 1968, Copyright American Meteorological Society).
dramatically, with values between 0 kt and 20 kt. There is also a favorable pattern set up in the NW Pacific. It is no coincidence that the Northern Hemisphere hurricane season peaks in August.

**TC Development Characteristics By Basin**

**Northeast Pacific**

Storms develop here from about late May through the month of October. This basin is unique because storms that form here do not recurve into the mid-latitude westerlies – usually they move toward the central Pacific, gradually spinning down as they move over cooler SSTs. Another scenario is that some move northward and make landfall over the Mexican or Central American coast.

**Northwest Pacific**

You may have noticed from the climatological wind shear maps that certain areas of this region have favorable vertical wind shear year round. In fact, tropical cyclogenesis does occur year round in the NW Pacific, but the vast majority form during the summertime. Development is strongly influenced by the location of the equatorial trough (a broad area of low-level convergence and convection forced by the general circulation of the Earth. Also referred to as the Intertropical Convergence Zone (ITCZ)).

About 1/3 of TC formations globally occur in this basin. The reason for this is a combination of the large area of warm SSTs and climatologically small vertical wind shear.

**North Indian Ocean**

Development in this region is strongly controlled by the progression of the ITCZ. During February through May, the ITCZ swings up north into the Bay of Bengal. TC genesis tends to follow this pattern northward. As the ITCZ retreats southward, so does TC genesis. Genesis can occur in all months except the winter months of late November through early February.

**South Indian Ocean**

This area experiences on average about 6 tropical cyclones in a year. There is a large area of development northeast of Madagascar, corresponding to the farthest pole ward progression of the equatorial trough. The central and Indian Ocean sees less development as the equatorial trough remains closer to the Equator during the summer months.

**Northwest Australia and South Pacific**

About 2 storms each year develop off the NW Australian coast (105°-135°E), 3 off the NE Australian coast (135°-150°E), and 4 in the South Pacific (150°E to 150°W). The large vertical wind shears present over much of the year in this area tends to limit development poleward of 20°S.
North Atlantic

This region is interesting for a number of reasons. First, TCs that form here directly impact the United States. Second, there is a distinct seasonal shift in development areas by month. At the beginning of the season (June and July), development is restricted to the western Caribbean by the still cool SSTs. During the months of August and September, development shifts eastward to the central and eastern Atlantic as SSTs warm and vertical wind shear over the basin decreases. Finally, the months of October and November show development returning to the western part of the basin. The main limiting factor to development in October and November is the southern advance of the mid-latitude westerlies, which increases the vertical wind shear over the basin. SSTs during those months are still well-above the critical threshold of 26.5°C needed for genesis.

The third interesting aspect of the North Atlantic is the origin of the disturbances that turn into TCs. A large portion of TCs form from easterly waves. These are troughs in the atmosphere spawned by a unique set of circumstances over the African continent. Easterly waves move toward the west at about 10-15 m/s and frequently have intense convection associated with them – a perfect pre-condition for TC formation. Easterly waves will be examined in more detail later in the course.

Finally, development in this region frequently occur pole ward of 20°N. This is the only basin where this happens.

Summary

A brief climatology of tropical storm formation was presented. Since Atlantic Ocean storms are the main area of concern for U.S. residents, the rest of the tropical cyclone discussion will focus on this area (although much of the discussion is relevant for storms anywhere on Earth).

Review Terms

Equatorial trough

Easterly waves

References

Chapter 3
Tropical Cyclone Structure

Since reconnaissance flights began into tropical cyclones in the 1940s, scientists have gained a great understanding of the structure of them. Most lay people are aware of the structural characteristics of TCs that can be seen from satellites: the eye, eyewall, and perhaps rainbands. But if we look inside (and under) the clouds, a rich and complicated picture of the TC emerges. Most of the information presented here was found in a study of research flights into four mature hurricanes: Anita (1977), David (1979), Frederic (1979), and Allen (1980). The results are summarized in Jorgensen (1984a, 1984b).

Basics of Tropical Cyclone Structure and Flows

The genesis and development of tropical cyclones will be discussed in detail in the following section. This section will provide some basics on the structure of a mature tropical cyclone, including visible features and wind flows.

Figure 4 is a gross representation of the major flows and features in a tropical cyclone. The cross section allows us to view the secondary circulation, which is the vertical circulation cell forced by low-level convergence and buoyancy in the eyewall region (the primary circulation refers to the horizontal winds caused by the pressure gradient force). Low-level (1-2 km) inflow evaporates moisture from the warm ocean surface in a process called isothermal expansion. Air expands adiabatically (and thus cools) as it approaches lower surface pressures in the core, but maintains its temperature through sensible heat fluxes from the warm ocean. In the hurricane core, air rises within cumulonimbi clouds and cools adiabatically. Latent heat is released in large quantities there, partially warming the TC core. At the top of the storm, the air flows out and loses energy through electromagnetic radiation to space. Finally, the air sinks back toward the surface (warming adiabatically) at some distance from the TC. This process describes the hurricane as a Carnot heat engine and is described in more detail in Emanuel (1991).
These flows create the three primary features of a mature TC: the eye, eyewall, and rainbands. The **eye** was once thought to be a calm, passive observer to the violence of the TC circulation but that view is quickly becoming modified. The eye of a TC is formed by the subsidence of air from near the tropopause level. The air is able to sink down to about the 850 mb level. The adiabatic warming of the subsidence (in addition to contributions from the latent heat release in the eyewall) creates a warm temperature anomaly in the upper troposphere within the eye. Figure 37 in section II (AMSU Floyd cross section) shows a cross section of temperature from an Advanced Microwave Sounding Unit (AMSU) satellite pass over Hurricane Floyd. Note that the anomaly is highest between 200 and 300 mb, with a magnitude greater than 10°C. Eyes have various sizes, ranging from 8 km to over 200 km across. The lower levels of the eye (close to the surface) are usually relatively moist, and commonly contain a lot of stratocumulus clouds. Surface winds in the eye may be rather strong near the eyewall boundary but generally decrease near the center. The lowest surface pressure of the TC occurs in the eye.

Figure 5 is a photo taken from a Hurricane Hunter aircraft. It was taken inside the eye of Hurricane Georges, probably at an altitude of at least 10,000 ft. Note that the lower portion of the eye appears to be overcast. There is a distinct boundary between the cloud tops and the clear air above – this is probably the level where the descending air from the tropopause stops descending.
The figure above (Figure 6) is a satellite image of the eye of Hurricane Isabel (2003). Here we can see features in the eye that suggest the region is dynamically active with motions forced by interactions with the surrounding eyewall. The black stars in the eye denote the centers of **mesovortices**, which are mesoscale circulations believed to be created by the eyewall. They rotate around the center of the eye like a BB rolling around in a rotating pipe. These features are currently being researched.

Figure 6 also reveals a sloping eyewall. The eyewall is the area of strongest surface winds, strongest updrafts, and highest cloud tops of the TC. Traditional theory is that mature TCs contain only one eyewall, but recent observations have shown that more than 50% of all TCs that
have surface winds of at least 120 kt have multiple eyewalls (concentric) present during some point of their lifecycle (Hawkins and Helveston 2004).

**Convective Structure of the Eyewall Region**

A more detailed picture of the eyewall region was found in the Jorgensen (1984a,1984b) study. Figure 7 shows a schematic of the wind and radar reflectivity from Hurricane Allen (1980). Allen was a category 5 hurricane that underwent multiple concentric eyewall cycles (see appendix A) during its lifetime. The intense banding signatures of precipitation in the eyewall region have been known for many years. However, one interesting revelation of the Jorgensen work is the rather light areas of precipitation that surround the intense convection, especially in more symmetric storms like Hurricane Allen.

Another key finding is that the tangential wind (circular wind around the center) peak is located about 2 km radially outward of the maximum vertical velocity. The radar reflectivity maximum is also outward from the updraft core. The eyewall of the hurricane slopes outward with height, with the smallest eye diameters close to the surface. For Hurricane Allen, the slope angle varied from 30° to 45°.

**Rainbands**

Jorgensen’s analysis found two types of rainbands: those that had a tangential wind maximum associated with them, and those that did not. He suggests that those rainbands with a positive tangential wind anomaly were actually nascent eye walls developing around the mature inner eyewall. This was observed in Hurricane Allen. The other surveyed storms had rainbands that were not accompanied by a bump in the tangential winds.

**Stratiform Precipitation Region**

Jorgensen found that over 90% of the rain areas in tropical cyclones were stratiform in type. That is, the precipitation was not the result of strong convective activity, as observed in the eyewall region. However, the convective rainfall contributes about 40% of the total storm precipitation (since convective rains are typically much more vigorous than stratiform rain). Also, data revealed the existence of a “bright band” in radar reflectivity just below the freezing level of the hurricane. This is the area where ice particles in the stratiform precipitation area are melting as they fall to the surface.
Figure 7. Schematic cross section depicting the locations of the clouds and precipitation, radius of maximum wind, and radial-vertical airflow through the eyewall of Hurricane Allen on 5 August 1980. Darker shaded regions denote the location of the largest radial and vertical velocity (From Jorgensen 1984b, Copyright American Meteorological Society).

**Review Terms**

Secondary circulation

Primary circulation

Isothermal expansion

Carnot heat engine

Eye

Eyewall
Mesovortices

References


Chapter 4  
Tropical Cyclone Lifecycle

**Tropical Cyclogenesis**

*Tropical cyclogenesis* (TCG) is the transformation of a group of disorganized thunderstorms into a self-sustaining synoptic-scale vortex. There are several theories on how this process occurs – they will be discussed later in this section. It is accepted that there are environmental conditions that are necessary (but not sufficient) for TCG to occur. Riehl (1948), Gray (1968), and McBride and Zehr (1981) define these as:

1. **Pre-existing convection** – Provides the latent heating to create the warm core
2. **Significant planetary vorticity** – Provides the spin to the system; otherwise would be just an area of low-level convergence. Generally, systems must be at least 2°-3° latitude away from the Equator, although TCG is rare within 5° of the Equator.
3. **Favorable wind shear pattern** – Must be small over the center of the developing system. Otherwise, latent heat will get carried away from the developing center and the system cannot produce more inflow at low-levels.
4. **Moist mid-troposphere** – A dry troposphere is death to a cloud cluster. It causes high evaporation (and thus cooling), snuffing out any chance of development.
5. **Warm ocean with a deep mixed layer** – Generally, sea surface temperatures (SSTs) must exceed 26.5°C. This temperature allows for a sufficient amount of evaporation by the inflowing air to sustain enough latent heat release in the core. The mixed layer is the ocean layer in which water temperatures do not decrease dramatically with depth. Deeper is better. As surface winds increase, they churn up the ocean and bring up water from depth. If this water is very cool, it can kill off the system.
6. **Conditionally unstable atmosphere** – Encourages convection

These conditions frequently exist during the late-summer/autumn seasons across much of the world’s oceans. But TCG is a relatively rare event. For example, the Atlantic Basin produces about 100 or so “cloud clusters” (incipient convective systems) in a given season. On average, about 15 of those clusters undergoes TCG. The following theories try to explain how those 15 are born, and why 85% of cloud clusters dissipate into oblivion.

*Convective Instability of the Second Kind (CISK)*

Charney and Eliassen (1964) asked the question “Why do cyclones form in a conditionally unstable tropical atmosphere whose vertical thermal structure is apparently more favorable to small-scale cumulus convection than to convective circulations of tropical cyclone scale?” Their solution lies in viewing the interaction of small cumulus clouds with the large-scale circulation as a cooperative one. This conceptual view of tropical cyclone formation became known as **conditional instability of the second kind**, or **CISK**. As summarized in McBride (1995), CISK theory makes three basic assumptions:
1) the initial perturbation is a synoptic-scale wave with balanced dynamics  
2) frictionally induced upward motion will result in latent heat release in the free atmosphere above the low-level cyclonic vorticity  
3) the magnitude of the latent heat release is proportional to Ekman pumping

In addition, the tropical atmosphere is assumed to be conditionally unstable. The essence of the theory is a positive feedback loop, where latent heat release caused by the large-scale circulation in turn reinforces it. Ekman pumping (vertical motion forced by horizontal frictional convergence) initiated from the large-scale vorticity field results in upward motion and latent heat release in the column. This forces the development of a secondary circulation and increased inward flow into the column. In turn, the vertical vortex is “stretched”, resulting in increased cyclonic vorticity at the surface and hence a greater amount of Ekman pumping.

Charney and Eliassen showed that a development period of approximately 2.5 days over a 100 km region is obtainable with reasonable choices of input parameters into their 2-level model. These values are similar to scales of tropical depression formation. This type of development was not obtained when regular conditional instability was assumed.

Wind Induced Surface Heat Exchange (WISHE)

Arguments refuting CISK and the convective parameterizations based on it have recently been suggested. Xu and Emanuel (1989) disputed the crucial assumption of CISK that the tropical atmosphere was generally conditionally unstable, presenting evidence that the atmosphere was in fact near neutral to moist convection. Without a reservoir of convective available potential energy (CAPE) to tap into, CISK could not exist. Another argument against CISK theory is the seemingly incorrect connection between latent heating and temperature made by Charney and Eliassen, who assumed that latent heating directly leads to kinetic energy production. Emanuel et al. (1994) argues that adiabatic cooling and radiative heat loss nearly offset the positive contribution of latent heat release, and that the correlation between heating and temperature is very difficult to determine.

Energy contribution by the ocean has long been recognized as a crucial component of TCG and maintenance (e.g. Riehl 1954). Seizing upon this and the weaknesses of CISK, Emanuel and others developed a new theory; ultimately named wind induced surface heat exchange, or WISHE (Emanuel 1989; Emanuel et al. 1994). A model based on WISHE can produce an amplifying tropical storm without the assumption of conditional instability.

Emanuel et al. (1994) presents the scenario of TCG within a WISHE framework. An incipient vortex induces Ekman pumping, resulting in upward motion throughout the depth of the troposphere (assuming a length-scale of 500 km). Eventually, downdrafts result from this forcing, bringing low θE air into the sub-cloud layer. Heat flux from the ocean surface initially is not enough to counteract this effect, and the vortex threatens to cool and spin down. The key factor that allows for amplification of the vortex is a moistening of the sub-cloud layer (through stratiform precipitation) to near saturation. Hence, when the moistened air is cycled into the secondary circulation, low θE air eventually disappears. This allows latent heat flux from the wind induced surface evaporation to begin to dominate, warming the core and allowing growth of the vortex. Thus, WISHE creates its own conditional instability through energy extraction from the ocean surface. An observational study conducted in Hurricane Guillermo (1991) during...
the Tropical Experiment in Mexico (TEXMEX) provides evidence for the processes hypothesized in WISHE (Bister and Emanuel 1997). Figure 8 is a schematic from that paper that illustrates the WISHE process.

Mesoscale Convective Vortices

A mesoscale convective system (MCS) is a large group ( > 100,000 km² in area) of organized convective clouds that persist for several hours. Within the MCS, there have been observations of the development of a localized area of enhanced cyclonic vorticity, called mesoscale convective vortices, or MCVs. They usually form in the rear stratiform precipitation region of the MCS.

MCVs commonly spawn severe weather over land areas. They are also theorized to lead to tropical cyclogenesis through a couple of ways. The first theory is based on what is called “vortex tube stretching”. The MCV may stretch due to downdrafts created by evaporative cooling in the stratiform precipitation region. This stretching will strengthen the circulation (analogous to a figure skater pulling in his/her arms while spinning) and, if it does reach the lower levels, begin to tap into the high energy air near the ocean surface. The second theory revolves around the mergers of two or more MCVs. The merger may result in one stronger MCV, which can then provide the basis for TCG. There is some observational evidence to support this hypothesis (Simpson et al. 1997).

Tropical Depression

When TCG occurs, the system is usually classified as a tropical depression (TD). There are usually one or two instances in a given season in which the system is initially classified as a tropical storm. A tropical depression must have a closed circulation. Forecasters look for evidence of a west wind south of the circulation (in the Northern Hemisphere). Until recently, they had to look for low-level, eastward moving clouds in geostationary imagery or be lucky enough to have the system move over a buoy or ship. In recent years the QuikSCAT satellite has provided timely information on wind speed and direction over developing systems, allowing forecasters to evaluate TD formations more accurately. The QuikSCAT satellite is discussed in more detail in the next section. Maximum sustained winds for tropical depressions cannot exceed 17.5 m/s (39 mph).

In the Atlantic basin, TDs are given a number. Numbers start at 1 with each season. In Australia, TDs are called “tropical lows”. India separates depressions into “Depression” (8.5 – 13.5 m/s) and “Deep Depression” (14 – 16.5 m/s).
Figure 8. Conceptual model of tropical cyclogenesis from a preexisting MCS. (a) Evaporation of stratiform precipitation cools and moistens the upper part of the lower troposphere; forced subsidence leads to warming and drying of the lower part. (b) After several hours there is a cold and relatively moist anomaly in the whole lower troposphere. (c) After some recovery of the boundary layer $\theta_e$ convection redevelops (From Bister and Emanuel 1997, Copyright American Meteorological Society).

TDs have a rather disorganized look to them from above. Figure 9 shows a tropical depression in the Gulf of Mexico. Note that there is little evidence of banding and the system is fairly asymmetric. However, there is a large area of deep convection and large amounts of latent heat being released into the atmosphere. Frequently, the circulation center at the surface is ill-defined and has a tendency to jump around with the intense convection. This makes TDs extremely difficult to forecast in terms of track. Figure 10 shows the “best track” (best estimate of storm location) image for TD-10 (2005). Note
that the center appears to change direction rather quickly in the early stages of the system. In reality, the center is probably reforming within a broader center of circulation. The track of TD-10 shown in Fig. 10 was probably drastically smoothed during post-storm analysis.

**Tropical Storm**

If maximum sustained winds in the TD exceed 39 mph, the system is upgraded to a tropical storm and given a name. Tropical cyclones had been given names in the West Indies for hundreds of years, usually after the saint’s day on which the storm occurred. Earlier in the 20th century, latitude and longitude coordinates were used, but this was quickly found to be confusing, especially when multiple storms existed at once. During World War II, it became common practice to use a woman’s name. This persisted until 1978, when men’s names were formally incorporated into the lists. Those lists are internationally agreed upon at the World Meteorological Organization. In the Atlantic basin, there are 6 separate lists of names that are rotated through. When a storm has been historic (either in terms of damage created, fatalities, or meteorological significance), the NHC proposes that the name be “retired”. If the WMO agrees, a new name is chosen to replace the retired name. There are no storms that begin with the letters Q,U,X,Y, and Z since names that begin with those letters are scarce.

Figure 11 shows a satellite view of Tropical Storm Katrina (2005). First thing to note is that the storm is much more symmetrical than the depression shown above. Also, the color enhancement allows us to see a large, overshooting top near the center of the circulation. This is a concentrated area of intense convection and a signal that the storm is in an intensification period. Banding features are now becoming evident, especially on the east side of the system. Also seen are high cirrus outflow in nearly all four quadrants (suppressed some on the north side – Katrina is probably experiencing some southerly wind shear at the time). The absence of an eye feature indicates that the storm is not of hurricane force but there are many features that suggest that this is a healthy tropical storm.

Tropical storms have maximum sustained winds between 40 and 73 mph. They are called “Cyclonic storms” or “Severe Cyclonic Storms” in India and “Tropical cyclones” in Australia.
Figure 10. Best-track for TD-10 (2005). Note the irregularity of the track over the 4 days.

Figure 11. Tropical Storm Katrina (2005). The center is probably under the convective burst at upper left.
**Hurricanes**

Once it is determined that the maximum sustained winds are 74 mph or more, the tropical storm is classified as a hurricane (“Very Severe Cyclonic Storm” in India, “Severe tropical cyclone” in Australia, “Typhoon” in the western Pacific). The structure and characteristics of hurricanes were presented in detail earlier. The Saffir-Simpson scale for classifying hurricanes is presented in the section on “Tropical Cyclone Impacts”.

**Dissipation**

There are several factors that will cause a tropical cyclone to weaken or dissipate:

1) Center of circulation moves over land
2) Storm circulation draws in dry air from land or other dry atmospheric layer
3) Storm moves over colder water
4) Storm remains stationary for too long
5) Negative atmospheric dynamics (e.g. increasing vertical wind shear)

Landfall causes the most dramatic weakening. A mature tropical cyclone will weaken to a TD or a remnant low usually within 48 hours of making landfall. The energy source (latent heat extracted from the warm ocean) to the core is cut off, leaving the vortex to a fate of frictional dissipation as now cooler and drier air arrive in the core. Also, it is quite common for TCs to weaken *as they approach* close to land. This was seen with Katrina during the 2005 season. It appeared that Katrina’s circulation ingested dry continental air as it moved towards the New Orleans coastline, which probably contributed to its weakening before landfall. Numerous other northward moving Gulf of Mexico storms have also exhibited this behavior (e.g. Ivan (2004)).

The movement over, or churning up of, cooler water will lead to weakening but not as quickly as what happens over land. There is still some energy taken into the system, though not enough to maintain its intensity. Furthermore, frictional effects over the ocean surface are much less severe. A stationary storm is likely to induce oceanic upwelling, creating cooler SSTs under the storm circulation. This effect is lessened if the mixed layer of the ocean is deep.

**Extratropical Transition**

A common scenario that plays out in all of the TC basins is the recurvature of TCs into the mid-latitudes. As TCs move into the mid-latitudes (and cooler waters), they experience fundamental changes to their thermodynamic structure. Namely, the warm core center disappears and temperature gradients (front-like features) begin to develop in association with the system. This process is called *extratropical transition*.

Recent research (Hart 2003) has provided a lot of insight into the phases of cyclones worldwide. It has been particularly helpful in determining whether a cyclone is tropical, sub-tropical, or extratropical and when it passes through each category. The “phase space” determines the cyclone phase by examining three measures of phase:
• **900 – 600 mb thickness gradient across the cyclone** – This is a measure of the temperature gradient across the cyclone. As shown by the hypsometric equation, thicker layers of atmosphere have higher mean temperatures in the layer. If the thickness is highest near the center of the cyclone (warm core) and decreases symmetrically outward, it is a good indication of a tropical cyclone.

• **900 – 600 mb vertical geopotential height gradient** – If the height gradient increases as you move vertically, that indicates a non-tropical (cold core) system. If the height gradient decreases in the vertical, a tropical system is diagnosed.

• **600 – 300 mb vertical geopotential height gradient** – Measures the cyclone phase for the upper troposphere, to help distinguish between shallow warm core systems (subtropical storms) from deep warm core systems (tropical cyclones).

These measures are calculated along the track of the tropical cyclone to help determine when extratropical transition will occur. It is graphically represented in a “phase space diagram” (Figure 12). The phase space diagram is separated into four regimes:

**Figure 12. Hart phase space diagram for Alberto (2006) around the time of it’s Florida landfall.**

symmetric warm core, asymmetric warm core, symmetric cold core, and asymmetric cold core. The letters “A”, “C”, and “Z” correspond to locations on the track map on the upper right. From the diagram, we can see that Alberto (2006) was unquestionably warm core but should begin to...
make its extratropical transition over the southeastern United States, completing its transition over the North Atlantic.

The phase space diagram is also commonly used by forecasters at NHC to determine the characteristics of mid-latitude cyclones (whether it is tropical, subtropical, or extratropical).

**Study Terms**

Tropical Cyclogenesis

Mixed layer

CISK

WISHE

MCS

MCV

Tropical Depression

Tropical Storm

Extratropical transition

Phase space diagram

**References**


In a reasonable thermodynamic and dynamic environment (e.g. low to moderate vertical wind shear, sufficient SSTs and planetary vorticity, moist troposphere), a tropical cyclone will tend to intensify up until it reaches a theoretical intensity limit. Of course, it will never be one slow, steady intensification trend. Intensity change in tropical cyclones is not well understood at all. There are influences that affect TC intensity at multiple scales: planetary-synoptic (environmental humidity, ocean SST/heat content), synoptic (local wind shear patterns), mesoscale (convective features, latent heating distributions, mesoscale vortex interactions, eyewall cycles), and microscale (air-sea interface, phase changes between vapor/water/ice) plus random interactions between scales.

This section will address a few of the major ideas in tropical cyclone intensity change. The first section will discuss the theoretical thermodynamic limit of tropical cyclone intensity mentioned above. This is followed by brief discussions on a couple of mechanisms that affect intensity (concentric eyewall cycles and trough interactions) and a look at rapid intensification.

**Maximum Potential Intensity (MPI)**

The intensification process of a TC occurs as long as there is a net mass loss in the column over the center of the surface circulation. The maximum potential intensity (MPI) is the theoretical upper limit of intensity that a TC can achieve. One of the first papers to appear on the topic was written by Banner Miller (1958). He proposed that “the minimum pressure that can occur within a hurricane is related to the temperature of the sea surface over which it moves”. The SST controls the temperature and moisture content of the air immediately above (air which is lifted in the storm) – lower the SST and the amount of energy available to the storm drops. Miller’s analysis produced a SST vs. maximum intensity curve, shown in Figure 79.

![Figure 79. Minimum probable pressure within a hurricane over various sea-surface temperatures (From Miller 1958, Copyright American Meteorological Society). For reference, 29.00” = 979 mb, 27.00” = 912 mb, 26.00” = 878 mb.](image)

Miller’s MPI also relies on the existence of Convective Available Potential Energy (CAPE) in the tropical atmosphere. CAPE provides an energy source for the air parcels that are assumed to be in moist adiabatic ascent. Recently there have been a couple of other formulations of MPI. Emanuel (1986, 1988) envisioned the TC as a Carnot heat engine, whereby the intensity of the cyclone is due to the
difference in the surface temperature (SST) and the temperature at the “outflow” level of the atmosphere. The idea of the Carnot heat engine is to transfer energy from a warm region (ocean surface and lower troposphere) to a cool region (outflow layer), converting potential energy into mechanical energy (updrafts, leading to inflow) in the process. Another recent MPI formulation was made by Holland (1998). Holland’s formulation is similar to Miller’s, in that it relies on the presence of CAPE to determine the MPI. However, Figure 80 shows that Holland’s method produces unrealistic minimum pressures (770 mb for an SST = 31°C) than Miller’s method. This is attributed to Holland’s attempts to incorporate the effects of the presence of an eye in the hurricane. Figure 81 shows a comparison of Emanuel’s MPI model with Holland’s.

If we just look at the intensities of tropical cyclones around the world and calculate the SST for each, we can derive an empirical MPI for SST. This was done by DeMaria and Kaplan (1994) for Atlantic basin storms (this was eventually included in their SHIPS model for intensity change prediction). Figure 82 shows the maximum wind speed by SST for their sample of Atlantic storms. Note that a 75 m/s wind corresponds to a Category-5 hurricane and lies closer to the Emanuel theoretical model in Fig. 82 than Holland’s.
Concentric Eyewall Cycles

Even the most conservative theoretical MPI formulations predict unrealistic intensities for very warm SSTs (Holland’s 770 mb hurricane, Emanuel “hypercanes” (see paper)), or at least intensities that are much lower than the observed values shown in Fig. 82. That result suggests
that there are processes occurring in real TCs that are limiting actual intensities to well below their thermodynamic theoretical limit. One such process is the **concentric eyewall cycle**.

A concentric eyewall is the situation when a band of heavy convection develops around the primary eyewall. The secondary band contains a secondary wind maximum and may act as a barrier to the flow of high $\theta_e$ air into the central vortex. This has the effect of initially weakening the TC. However, usually the outer ring will begin to contract and eventually take the place of the inner eyewall. Intensification, sometimes rapid intensification, occurs during this phase. Concentric eyewall cycles are common in very strong tropical cyclones. Individual storms may undergo several cycles during their lifetime.

Black and Willoughby (1992) describe the evolution of eyewall cycles in Hurricane Gilbert (1988), which until 2005 (Wilma) was the strongest hurricane ever observed in the Atlantic basin at 888 mb. Figure 83 shows the classical satellite presentation of a concentric eyewall cycle at the point of the outer eyewall completely surrounding the inner eyewall. The TC has the appearance of a “moat” around its center. The moat is characterized by a localized wind minimum, compared to the two rings on either side.

The moat in this figure has a radius of about 25 km, with a very small pinhole eye in the center (~7km). A series of aircraft flights through Gilbert revealed a classical double wind maxima structure (Figure 84). The middle panel clearly shows the primary eyewall (I), the wind minima outside, and then the secondary maxima between 50 km and 100km from the center of the storm.

Figure 83. Radar reflectivity composite from 0923-1126 UTC

*on 14-September 1988 (From Black and Willoughby 1992, Copyright American Meteorological Society).*
Figure 84. Flight-level tangential wind speed from south-north traverses through the center of Hurricane Gilbert. Bold I’s and O’s denote the location of the inner- and outer-eyewall wind maxima, respectively. Times at the beginning and end of each radial pass are plotted at the top of the panels (from Black and Willoughby 1992, Copyright American Meteorological Society).

**Trough Interactions**

Soon after Hurricane Opal (1995) rapidly intensified in the Gulf of Mexico on its way to a Florida panhandle landfall, some researchers attributed the intensification period to a **trough interaction**. A trough interaction occurs when a TC encounters an **eddy flux convergence (EFC)** of wind $>10$ (m/s) per day. The EFC is the amount of convergence at upper levels (200 mb) due to eddies in the wind field, and is defined mathematically as:

$$EFC = -\frac{1}{r^2} \frac{\partial}{\partial r} r^2 u'_r v'_l$$
where \( u \) is the radial wind, \( v \) is the azimuthal wind, \( r \) is the distance from the storm center, and the primes indicate deviations from the azimuthal mean. The ‘L’ refers to storm-relative flow. A high EFC indicates the existence of an upper-level trough in the vicinity.

Hanley et al. (2001) looked at all TCs in the Atlantic during the period 1985-1996. They found the following:

- Trough interactions occurred about 25% of the time
- TCs over warm water and away from land intensified 78% of the time when a trough was directly superpositioned over the system (61% intensification when the trough was farther away)
- The relative scales of the trough and TC were important – if the trough was too large, wind shear would increase too much and the effect on TCs would be negative.

Hanley et al. note that the intensification sign of the TC/trough interaction depends on rather subtle changes in the characteristics of the trough, making it difficult to forecast such an interaction correctly.

**Saharan Air Layer (SAL) Interactions**

It has been known since at least the early 1970s that massive dry air intrusions occur over the Atlantic basin during the late spring and early fall. These outbreaks of westward moving dry air from the African continent (called the Saharan Air Layer, or SAL) can have profound impacts on tropical cyclone intensification. The dry air can be seen via GOES satellite imagery by a technique developed by Dunion and Velden (2004). This technique will not be presented in detail here, but you are encouraged to find the paper if you want to learn more. Basically, they analyze the brightness temperature \((T_b)\) differences between two GOES-12 channels (3.9 and 10.7 \(\mu\)m). A “normal” (non-SAL) difference in \(\Delta T_b\) (10.7 minus 3.9) will 6°-10°C, while SALs will show differences on the order of -13°-5°C. Figure ? shows a satellite image of a SAL outbreak over the eastern Atlantic.

**SAL Characteristics**

The SAL originates as a deep pool of dry air extending from the surface up to approximately 500 mb over the summer African continent. Easterly waves may pick up the dry air and move it along with the wave axis toward the Atlantic. As the wave approaches the Atlantic, a moist marine layer modifies the lower boundary of the dry air, producing the SAL. As the SAL moves offshore, the lower (upper) boundary of dry air can be found about 1-2 km (5.5 km) above the surface. Amazingly, the SAL can maintain its warm, stable thermodynamic structure across the entire Atlantic basin. It’s large size (some outbreaks are about the same size as the contiguous United States) and extreme conditions create significant impacts for the tropical Atlantic and potential tropical cyclones in the basin.
Figure ?. A SAL outbreak as shown by a SeaWIFS satellite in February of 2000. The Saharan dust can tranverse the entire Atlantic, turning the sky red along the eastern U.S.

SAL Interactions with Tropical Cyclones

Although convection can be enhanced around the southern and western boundaries of the SAL (most likely due to the strong moisture gradient found there), the interior portions present a hostile environment for developing and mature TCs. The dry air works to evaporate developing convective towers, creating an inhibiting cooling effect due to the latent heat of evaporation. In addition, an atmosphere containing a SAL is very stable - warm air aloft sits upon cooler, denser marine air below. Figure ? shows differences between SAL and non-SAL moisture soundings.

When an easterly wave or developed TC becomes enveloped by the SAL, deep convection tends to dissipate, sometimes quite rapidly. Thus, it is possible to significantly weaken or even kill mature TCs in a climatologically favorable place and time (warm SSTs, little or no shear, conditionally unstable environment, etc.). Before the ability to identify and track the SAL were made, forecasters were frustrated by many cases where a TC failed to intensify or develop in a very favorable (at least to their eyes) environment.
The SAL layer(s) are tracked in real time at:

http://cimss.ssec.wisc.edu/tropic/real-time/wavtrak/sal.html

**Rapid Intensification (RI)**

**Rapid intensification (RI)** is the explosive deepening of a tropical cyclone. Kaplan and DeMaria (2003) define RI as a maximum sustained surface wind speed increase of 15.4 m/s (30 kt) over a 24-hour period. This corresponded to the 95th percentile of all 24-hour intensity changes in the Atlantic basin from 1989-2000 (basically the top 5% of intensity changes). *All* category 4 and 5 hurricanes, 83% of all major hurricanes (category 3, 4, or 5), 60% of all hurricanes, and 31% of all tropical cyclones experienced at least one RI period during their lifetime.

Kaplan and DeMaria (2003) performed a statistical analysis on RI Atlantic storms from 1989-2000. Figure 85 shows the tracks from the Atlantic storms during their RI phase. Note that most occur south of 30°N and that their appears to be a lack of RI cases in the eastern Gulf of Mexico and eastern Caribbean. Also shown are the months in which RI occurred. About 72% of all cases occurred in August and September, with more later in the season (Oct. and Nov.) than earlier (June and July).

Other main conclusions from Kaplan and DeMaria:

- RI cases were farther from their MPI and occurred in areas of higher SST and higher lower-tropospheric relative humidity
• RI was more likely in systems that were not impacted by upper-level features such as troughs or cold lows. Some have suggested that the RI of Hurricane Opal (1995) was due to a trough interaction (more on Opal in the next section).

• RI probability can be estimated through the analysis of five predictors: previous 12-hour intensity change (already deepening storms more likely), SST (higher more likely), low-level relative humidity (higher more likely), vertical shear (lower is better), and difference between current intensity and MPI (larger is better).

Kaplan and DeMaria’s work is incorporated into a “RI Probability” in the SHIPS hurricane intensity model. Their work has been a valuable contribution in the goal of predicting RI in a forecast environment.

Figure 85. The 24-h tracks of the 1989-2000 RI cases. The distribution of the RI cases by the month when they occurred is also presented (From Kaplan and DeMaria 2003, Copyright American Meteorological Society).

Hurricane Opal (1995) and Warm Core Eddies

During the early morning hours of October 4, 1995, residents along the northern gulf coast slept as a once dormant hurricane suddenly accelerated toward them, intensifying into a monster category 4 hurricane in just a few hours (pressure dropped from 963 mb to 916 mb). From 0900 Z to 1100 Z, the central pressure of Opal plummeted from 933 mb to 916 mb. As Opal continued to accelerate and intensify, forecasters at NHC scrambled to contact emergency management officials to get hurried evacuations rolling. Fortunately, by the time Opal made landfall 11 hours later, its pressure had risen to 940 mb and winds had subsided to weak category 3 status. Forecasters asked themselves what caused the RI and subsequent rapid weakening.
Opal is a great example of the interplay of factors that cause TC intensification change. The post-storm report by NHC (http://www.nhc.noaa.gov/1995opal.html) attributed the RI to favorable environmental conditions (28°C - 29°C water), a well-established anti-cyclone over the Gulf of Mexico, and perhaps the termination of a concentric eyewall cycle that resulted in a small (10 nmi) eye near the time of maximum intensity. Post-storm research by others (e.g. Shay et al. 2000, Hong et al. 2000) suggest that a **warm core ring**, or **warm core eddy** in the Gulf of Mexico was the primary forcing mechanism that caused the intensification (and subsequent weakening as Opal left the area).

A warm core eddy (or ring) is an ocean circulation of much higher than normal SST and oceanic heat content. In the Gulf of Mexico, warm core eddies originate from perturbations in the “**Loop Current**”, a warm ocean current that flows through the Florida straits and merges with the Gulf Stream. Frequently, the loop current will break and spin off a large area of warm water with a deep mixed layer. This eddy will spin around in the Gulf of Mexico for a number of weeks before being modified and dissipated. Warm core eddies can be thought of as nitrous fuel for hurricanes. They are a source of tremendous potential energy. When a TC traverses an eddy, it will undoubtedly intensify very quickly (unless the atmospheric factors are unfavorable) as latent and sensible heat fluxes increase suddenly and dramatically.

Figure 86 is from Shay et al. (2000). Note that Opal’s RI phase began soon after it moved over the warm core ring in the Gulf of Mexico. Conversely, the rapid weakening phase began soon after Opal left the warm core ring.

Other papers on Opal’s intensification (e.g. Bosart et al. 2000) attribute the RI of Opal to a trough interaction (shown to the northwest of Opal in Fig. 86). In reality, it was probably a “perfect storm” of favorable conditions that gave residents along the gulf coast a big scare that morning in 1995.

Figure 86. Track of Hurricane Opal (1995) in relation to the location of a warm core ring and an approaching upper level trough (From Shay et al. 2000, Copyright American Meteorological Society).

**Convective Bursts**
From the discussion of MPI and warm core eddies, it should be apparent that the ocean plays a critical role in TC intensification. There is an important coupling between the ocean and atmosphere that engenders a transfer of energy from low-levels into the upper troposphere. There is perhaps no more spectacular demonstration of that process than a **convective burst**. A convective burst, as defined by Rodgers et al. (2000), is “a mesoscale (100 km by hours) cloud system consisting of a cluster of high cumulonimbus towers within the inner core region [of a tropical cyclone] that approaches or reaches the tropopause with nearly undiluted cores.” The key term of this definition is *undiluted cores*. This means that the convective tower does not entrain outside air (which is typically cooler, drier, and more stable) into its updraft. The burst provides a pristine conduit for rapidly transferring latent and sensible heat into the upper troposphere. This enhances the warm-core anomaly aloft, accelerating upper-level divergence and spawning surface pressure falls.

Convective bursts are impressive events on satellite imagery. Figure 87 shows several examples of convective bursts. They appear as large, circular, continuous areas of extremely cold (-80°C - -100°C) cloud tops. Within the burst, there may be localized

![Figure 87. Examples of convective bursts. From Hennon (2006).](image)
Figure 88. Overview of burst chain of events: enhanced surface convergence and/or sea-air fluxes produce a favorable environment for convective burst occurrence. Sustained upper tropospheric energy release occurs, and subsiding air leads to an upper-level warm anomaly (From Hennon 2006).

overshooting tops – these have been called “hot towers” or “chimney clouds”, and have been the subject of intense research in recent years. Figure 88 illustrates the chain of events in a burst event.

A global survey of convective burst events from 1999-2001 was undertaken by Hennon (2006). Table 4 presents the convective bursts by ocean basin. Convective bursts are ubiquitous events globally, with 4 out of every 5 storms experiencing at least one burst.
during its lifetime. When a burst does occur, the TC will subsequently intensify 70% of the time. Figure 89 shows the change in wind speed following each of the 344 burst events documented by Hennon. There are very few cases when the TC weakened after experiencing a burst. For those weakening cases, most were undergoing extratropical transition (moving over cooler water).

### Table 4. Convective bursts by ocean basins (From Hennon 2006).

<table>
<thead>
<tr>
<th></th>
<th>IO</th>
<th>SH</th>
<th>WPAC</th>
<th>ATL</th>
<th>EPAC</th>
<th>GLOBE</th>
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<td># storms</td>
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<td>62</td>
<td>75</td>
<td>37</td>
<td>42</td>
<td>223</td>
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<tr>
<td># CBs</td>
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<td>107</td>
<td>106</td>
<td>60</td>
<td>56</td>
<td>344</td>
</tr>
<tr>
<td>ratio</td>
<td>1.5</td>
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<td>1.4</td>
<td>1.6</td>
<td>1.3</td>
<td>1.5</td>
</tr>
<tr>
<td>% storms with CB</td>
<td>100%</td>
<td>81%</td>
<td>76%</td>
<td>73%</td>
<td>90%</td>
<td>80%</td>
</tr>
<tr>
<td>% storms intensify</td>
<td>100%</td>
<td>79%</td>
<td>76%</td>
<td>73%</td>
<td>85%</td>
<td>79%</td>
</tr>
</tbody>
</table>

Figure 89. For \( n = 344 \) burst events (1999-2001), frequency distribution of intensity change (wind speed change in kt.) associated with convective bursts (From Hennon 2006).
**Review Terms**

Maximum potential intensity (MPI)

Concentric eyewall cycle

Trough interaction

Eddy flux convergence

SAL

Rapid intensification

Warm core eddy (ring)

Loop current

Convective burst

**References**


Chapter 6
Tropical Cyclone Forecasting

Forecasting tropical cyclones is a unique and challenging exercise. Unique because several hours are spent in determining the best forecast for a single synoptic-scale disturbance. Challenging for a variety of reasons:

1) Much is be learned about the physics of tropical cyclones
2) Since tropical cyclones are generally over water, little if any in situ data is available for the systems
3) Forecasts have direct consequences (e.g. evacuations, preparation) – a blown forecast could cost millions of dollars and even worse, lives lost

This section will focus on forecasting for the Atlantic Basin for two reasons. First, the author has experience in producing Atlantic Basin forecasts. Second, the Atlantic obviously produces all of the tropical cyclones that directly impact the United States.

Observation Sources

Chapter 8 will describe some of the observation sources that are routinely available to forecasters. It should be emphasized that the most critical observation source is the Geostationary Orbiting Environmental Satellite (GOES) array. These satellites provide a continuous view of the Atlantic and Pacific basins. Prior to the GOES era (early 1960s), forecasters had to rely on word of mouth, telephone, news reports, ship reports, etc. to determine where storms were located. Virtually the entire planet is now covered with geostationary satellites. Europe (METEOSAT), India (Metsat), and Japan (GMS) each operate its own constellations of geostationary satellites.

Model Guidance

In general, there are two primary types of computer models. Statistical models produce forecasts based on past events. Several variables are identified that are correlated with past outcomes. These variables are then evaluated for the current system. Statistical models can be an effective forecasting tool, especially for processes (e.g. tropical cyclone intensity change) that are not well understood physically. However, they are unable to predict extreme events with any accuracy since those occur quite rarely in the historical record. Dynamical models solve mathematical relationships given a set of input data. These are called “full physics models”, in that they rely on known physical laws (e.g. Equation of State, 1st Law of Thermodynamics, equations of motion, etc.) to produce forecasts. Unlike statistical models, dynamical models can in theory predict extreme events. However, they suffer when there isn’t a sufficient understanding of the physics or the input data is inaccurate or sparse. There is also a hybrid breed of models, sometimes called Statistical-Dynamical models, which are simpler than full-blown dynamical models in that they substitute less-understood physical processes with statistical relationships.
Track Models

A more detailed description of track models used at the National Hurricane Center can be found at http://www.srh.noaa.gov/ssd/nwpmodel/html/nhcmodel.htm (last updated in 2004). A brief summary of the track models, which are used to forecast where the system is going to go, is described below.

**CLIPER (CLImatology and PERsistence)** – A pure statistical track model. Analyzes historical track data from 1931-1970 to produce a track forecast at 12-hour intervals out to 120 hours. Generally has the least amount of skill. Other model forecasts are typically compared to CLIPER to evaluate forecast skill.

**NHC98** – A statistical-dynamical model. The latest in a long line of these types of models developed at NHC (NHC67, NHC72, NHC73, NHC83, NHC90). Different sets of equations are used based on storm location. For example, storms embedded in the easterlies tend to move to the right of the steering flow, while storms in the westerlies tend to move to the left. NHC98 attempts to take advantage of these observations. The actual forecast is a combination of CLIPER, observed mean geopotential heights, and forecast mean geopotential heights. The forecast heights are from the Global Forecast System (GFS) model – thus, NHC98 combines dynamical model output with statistics to make its forecasts. In reality, NHC98 is rarely used by forecasters at NHC.

**BAM (Beta Advection Model)** – Another hybrid model. Averages the steering wind through a certain depth of the atmosphere, then adjusts the track of the tropical cyclone based on the so-called beta effect (It has been shown that storms in zero steering flow will move the northwest just due to the change in the Coriolis parameter across the extent of the storm). There are three versions of this model, each using a different averaging layer: shallow (850-700 mb), medium (850-400 mb), and deep (850-200 mb). Although frequently consulted by forecasters, BAM models generally do not perform too well.

**LBAR (BARototropic model)** – A simple dynamical model. This is what is called a nested model. The model is initialized with three nested domains: 1) synoptic domain (entire basin), initialized with GFS analysis, 2) storm environment (50° box surrounding storm), initialized with any observations within that box and GFS as a background field, and 3) vortex environment (within 800 km of storm center), initialized with storm characteristics (intensity, size, motion). The LBAR can produce some skillful forecasts and runs very quickly. However, it struggles when the environment is complex (e.g. strong vertical wind shear or interacting storms).

Figure 13 shows the skill of each statistical/dynamical model described above relative to CLIPER for the 2003 Atlantic season. For this particular season, the BAM-Medium model was the most skillful, and NHC98 was generally the least skillful. However, performance varies widely from year to year (not shown).
Figure 13. Skill comparison of statistical/dynamical models at NHC for the 2003 hurricane season. Note that this cannot be extrapolated to other seasons (certain models will perform differently depending on the types of storms that occur in a given season). From http://www.srh.noaa.gov/ssd/nwpmodel/images/Skill03.gif

Figure 14. Comparison of track forecast errors during the 1995-1998 seasons for the GUNS consensus and its 3-model components.
**GFDL** (Geophysical Fluid Dynamics Laboratory model) – A full-physics model that predicts both track and intensity. It is coupled to the sea surface, meaning that it can recognize sea surface temperature changes as the model integrates. For the 2003-2004 seasons, the GFDL model had the best performance (for the 48-hr forecast) for any single model (see [http://www.nhc.noaa.gov/verification/figs/Early_model_ATL_trk_error_trend.gif](http://www.nhc.noaa.gov/verification/figs/Early_model_ATL_trk_error_trend.gif) for a figure).

**GFS** (Global Forecast System) – A full-physics model developed by NCEP for general use. Applied to tropical cyclones, the GFS is usually an above-average track model, especially compared to other non-dynamical models. For the 2003-2005 seasons, the GFS was the second-best single model behind the GFDL for the 48-hour track forecasts.

There are a few other dynamical models that you can research but will not be talked about in detail here. **NOGAPS** (Navy Operational Global Atmospheric Prediction System) and **UKMET** (United Kingdom Met Office Model) are two models that are used in consensus forecasts but generally perform more poorly than GFDL or GFS. **Consensus Models** (GUNS, GUNA, CONU) – A consensus model is quite simply an average of a number of models. In terms of hurricane prediction, this produces a 20% increase in track accuracy on average. Consensus models are frequently beat when it comes to a single forecast, but over the course of the season they will perform better than any one model. **GUNS** is the average of the GFDL, UKMET, and NOGAPS models. **GUNA** is the average of the GFDL, UKMET, NOGAPS, and GFS (formally the AVN). **CONU** is computed when at least two of the five possible consensus members are available – it doesn’t matter which two. All consensus products show very good track forecast skill. Figure 14 shows the average track errors for the GUNS model (purple) vs. its member components. These models are heavily relied upon at the NHC for track forecasts. See Goerss (2000) for more information.

**Intensity Models**

**SHIFOR** (Statistical Hurricane Intensity FORecast) – Analogous to the CLIPER track model; a purely statistical intensity model based on storms from 1900-1972. The predictors of SHIFOR include julian day (1 = 1 January, 365 = 31 December), initial intensity, intensity trend (change in last 12 hours), and latitude. Just as CLIPER, is considered the “no skill” intensity model and is thus used to grade other intensity forecasts.

**SHIPS** (Statistical Hurricane Intensity Prediction Scheme) – A statistical-dynamic hybrid intensity model (DeMaria and Kaplan 1999, DeMaria et al. 2005). Uses simple regression techniques with a variety of statistical, persistence, and synoptic predictors (including some GFS model-derived fields). Those predictors are summarized in a table from DeMaria et al. (2005), shown below.
Table 1. Predictors used in SHIPS 1997-2003. The predictors that are evaluated at the beginning of the forecast period are static (S), and predictors that are evaluated along the forecasted track of the storm are time dependent (T). An X indicates that the predictor was used in that year and a dash (--) indicates it was not used that year (From DeMaria et al. 2005, Copyright American Meteorological Society).

<table>
<thead>
<tr>
<th>Predictor</th>
<th>Static (S) or time dependent (T)</th>
<th>1997</th>
<th>1998</th>
<th>1999</th>
<th>2000</th>
<th>2001</th>
<th>2002</th>
<th>2003</th>
</tr>
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<tr>
<td>1) Absolute value of (Julian day – peak season value)</td>
<td>S</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>2) Gaussian function of (Julian day – peak value)</td>
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<td></td>
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<td>X</td>
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<tr>
<td>3) Initial maximum winds</td>
<td>S</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>4) Max wind change during the past 12 h</td>
<td>S</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>5) Initial max winds times previous 12-h change</td>
<td>S</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td>X</td>
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<tr>
<td>6) Zonal component of storm motion</td>
<td>S</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<td>7) Pressure level of storm steering</td>
<td>S</td>
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<td>X</td>
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<td>8) 200-hPa divergence</td>
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<td>X</td>
<td>X</td>
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<td>9) 200-hPa eddy momentum flux convergence</td>
<td>S</td>
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<td>10) 200-hPa eddy momentum flux convergence</td>
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<td>X</td>
<td>X</td>
<td>X</td>
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<td>11) Max potential intensity – current intensity</td>
<td>T</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<td>12) 850-200-hPa vertical shear</td>
<td>T</td>
<td>X</td>
<td>X</td>
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<td>13) 200-hPa zonal wind</td>
<td>T</td>
<td>X</td>
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<td>X</td>
<td>X</td>
<td>X</td>
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<td>14) 200-hPa temperature</td>
<td>T</td>
<td>X</td>
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<td>15) 850-700-hPa relative humidity</td>
<td>T</td>
<td>X</td>
<td>X</td>
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<td>X</td>
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<td>16) 500-300-hPa relative humidity</td>
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<td>X</td>
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<tr>
<td>17) 850-hPa relative vorticity</td>
<td>T</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<td>18) Surface – 200-hPa g, deviation of lifted parcel</td>
<td>T</td>
<td></td>
<td></td>
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<td>X</td>
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<tr>
<td>19) Vertical shear times sine of storm latitude</td>
<td>T</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>20) Square of potential – current intensity</td>
<td>T</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>21) Initial intensity time shear</td>
<td>T</td>
<td></td>
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</table>

A more recent addition to SHIPS is the inclusion of satellite data (GOES winds, ocean heat content). These have shown to improve intensity forecasts by about 7% through 72-hours. In general, SHIPS is a skillful intensity model and is relied upon very heavily by forecasters. There are few alternatives.

**GFDL** – As mentioned previously, the GFDL model also produces an intensity forecast. Usually, the GFDL is a bit more aggressive in its intensity forecasts than SHIPS (overly so many times). Both GFDL and SHIPS are examined and considered before an intensity forecast is made. Although SHIPS is frequently favored, the GFDL intensity forecast does impact the forecast thinking quite often.

**FSU Superensemble** – This is an experimental model developed at Florida State University that produces both a track and intensity forecast. It uses a “weighted consensus” approach. That is, it analyzes past biases in forecasts (including the official NHC forecast) and weights the errors. Its intensity skill is comparable to SHIPS.

Figure 15 shows the verification of the intensity models for the 2005 Atlantic Basin season. OFCL is the official NHC forecast, GFDL is the GFDL, DSHP is the SHIPS model, and FSSE is the Florida State Super ensemble. Note that DSHP and FSSE are close at all times, although no one model can beat the official NHC forecast.
Figure 15. Homogenous comparison for selected Atlantic basin early intensity guidance models for 2005, for pre-landfall verifications only. (From http://www.nhc.noaa.gov/verification/pdfs/Verification_2005.pdf, James Franklin)

Forecasting Tropical Cyclogenesis

Chapter 4 discussed in some detail the theories of tropical cyclogenesis (birth of a tropical depression) and the conditions that must be present in order for it to occur. Traditional dynamical models, like those discussed above in the context of track and intensity forecasting, have historically struggled to accurately predict tropical cyclogenesis. For example, the GFS had a nasty habit of spinning up so-called “boguscanes” all over the basin during the late 1990’s and early 2000’s. However, there are some indications that some dynamical models are improving in their prediction of tropical cyclogenesis. In early July of the 2008 Atlantic season, the NHC noted that Hurricane Bertha (a record holder for the farthest east a major hurricane had ever formed so early in the season, by the way) was correctly forecast by many global models, including the GFS. The discussion read:

ZCZC MIATCDAT2 ALL
TTAA00 KNHC DDHHMM
TROPICAL DEPRESSION TWO DISCUSSION NUMBER 1
NWS TPC/NATIONAL HURRICANE CENTER MIAMI FL AL022008
500 AM EDT THU JUL 03 2008

..... IT SHOULD BE NOTED THAT MOST GLOBAL MODELS...ESPECIALLY THE
GFS...SUGGESTED THE POSSIBILITY OF GENESIS IN THIS AREA OVER A WEEK
AGO...A REMARKABLE ACHIEVEMENT.
Whether this was a one-time success or a sign that global dynamical models now have skill in forecasting tropical cyclogenesis remains to be seen.

So why have dynamical models failed to have much skill in forecasting tropical cyclogenesis? Though there are probably too many reasons to list here, it is generally believed that there are three (3) major factors:

1. Insufficient observations
2. Insufficient model resolution
3. Lack of understanding of the physics of tropical cyclogenesis

As discussed in Chapter 4, there are a few different theories on how tropical cyclogenesis occurs. Without a clear resolution of #3, it is impossible to tell the model how to process the smaller-scale physics that occur within active convective seedlings. This is analogous to telling a pole vaulter to clear a 15’ bar with a 2’ twig.

Even if a clear consensus of genesis is determined, dynamical models will still struggle with genesis forecasting as long as #1 and #2 are not improved. Model resolution is continually increasing, now allowing for some inner core processes to be modeled explicitly. The number of observations is also on the rise, particularly from satellites. It should not be too long until we see consistently skillful tropical cyclogenesis forecasts emerge from the global dynamical models.

a. Statistical Schemes for Forecasting Tropical Cyclogenesis (through 48 hours)

During the “boguscane” days of genesis prediction there was (and still is) a need for skillful tropical cyclogenesis forecasts – this need was filled by statistical forecast schemes that relied primarily on the large-scale environmental conditions to assign a genesis probability. Two schemes will be briefly presented: the TC Formation probability product (developed by the Regional and Mesoscale Meteorology Branch (RAMMB) at Colorado State) and the Hennon scheme (yes, my graduate work).

The RAMMB product (see [http://www.ssd.noaa.gov/PS/TROP/genesis_description.html](http://www.ssd.noaa.gov/PS/TROP/genesis_description.html) for a description, DeMaria et al. 2001 for background) calculates a number of input parameters based off of GOES imagery and NCEP global analyses. It forecasts the probability of tropical cyclogenesis within a 5° x 5° box over the following 24 hours. It accomplishes this through a statistical technique called discriminant analysis, which basically looks at thousands of cases that already occurred to determine what conditions tend to lead to genesis and which do not. A “forecast function” is created from this “training” information. Then a new, independent set of data is fed into the equation and the routine calculates the probability that this new data will lead to tropical cyclogenesis. Figure ? shows the genesis probabilities for the 24-hour period beginning 12Z July 8, 2008. Note that current tropical cyclones are shown (which we don’t really concern ourselves with in this context). Note also that most probabilities are very low (< 4%) – rarely will the scheme show probabilities that exceed 15-20%.
Figure ?. Tropical cyclogenesis probabilities as calculated by the RAMMB statistical model. At this point, genesis is more favored in the Eastern Pacific than the entire Atlantic basin.

The six (6) parameters that go into the RAMMB model are: climatological formation probability, 850 hPa circulation, GOES cold pixel count, distance to a pre-existing TC, vertical shear, and percent of area over land.

Hennon and Hobgood (2003, hereafter HH) took a different approach. Instead of calculating probabilities for 5° boxes, they focused on the pre-existing convection (“cloud clusters”). This eliminated the need to calculate probabilities for vast areas of the ocean where genesis could not occur, since there were no large areas of deep convection to begin with. Their scheme is Lagrangian, meaning that it follows the cloud cluster as it moves around the basin.

HH also employed discriminant analysis to yield a probability of genesis. The selection of predictors was based off of the large-scale factors discussed near the beginning of chapter 4. They predictors were latitude, the daily genesis potential (DGP), maximum potential intensity, low-level moisture convergence, precipitable water, the 24-hour pressure tendency, and the 6-hour surface and mid-level vorticity tendencies. The DGP is simply the difference between the low (850 mb) and upper (200 mb) relative vorticity. It has been shown to be a strong predictor of tropical cyclogenesis.

HH showed significant skill in forecasting tropical cyclogenesis. To illustrate this, a case study will be presented. Figure ? shows the track of a cloud cluster that would eventually form into TD-15 (2000) and then Hurricane Keith (2000). Note the absence of cluster data points between west of 50°W and east of 80°W. The intense convection from the wave dissipated and was unable to be tracked; thus no predictors could be calculated.
Figure 2. Track of a cloud cluster that would eventually develop into TD-15 (2000). Each dot is a 6-hour time interval. From Hennon and Hobgood (2003).

Figure 2 shows the probabilities of genesis for each 6-hour period for the entire lifetime of the cloud cluster (up until genesis at 18 UTC September 28). Note that probabilities were not favorable (shaded region) for the first 2/3 of the cluster lifetime. Then, as convection re-emerged in the western Caribbean, probabilities were produced that ranged from 0.7 (70%) to 0.99 (99%). Keith would form into TD-15 a short time later.

Figure 3. Output probabilities (0 to 1, 1 meaning genesis is 100% likely) for the cloud cluster that developed into TD-15 on September 28, 2000 at 1800 UTC. The shaded region represents the regime where genesis would be unlikely. Each line represents a 48-hour forecast at 6-hour intervals. From Hennon and Hobgood (2003).

b. Long-range Tropical Cyclogenesis Forecasting (up to 30 days)
There has been some interesting work in the longer range tropical cyclogenesis forecasting area. This scheme, developed by Paul Roundy (Frank and Roundy 2006), looks at longer term factors in genesis such as the phase of the El Niño-Southern Oscillation (ENSO), existence of Madden-Julian Oscillation (MJO) forcings, and other atmospheric waves such as easterly waves, Kelvin waves, and Rossby waves. Many of these features will be discussed later in this text. But for now, you should be aware that these factors can have a significant influence on the likelihood of genesis and can be predicted out to at least a number of weeks and, as in the case of ENSO, sometimes longer. As one example, the passage of the active phase of the MJO will trigger enhanced convection in a region. This increases the probability of genesis, since strong convection will moisten the atmosphere and possibly destabilize it due to large amounts of latent heat release.

You are encouraged to browse Roundy’s web page, where he graphically presents his 30-day genesis forecast. Figure ? shows a 30-day forecast using his scheme. The probabilities can be interpreted as the likelihood that a tropical cyclone will exist in a particular region at the verification time. So, Roundy’s system is not a tropical cyclogenesis prediction scheme per se, but rather an assessment of the long-range factors that have shown to be connected to conditions favorable for TCs. It is a young area of research that will probably yield some very useful results in the near future. Roundy’s website is http://www.atmos.albany.edu/facstaff/roundy/tcforecast/tcforecast.html.

NHC Forecast Performance
Figure 16 shows the trend in official forecast track errors from 1970-2005. Errors have dropped dramatically for all forecast times since 1970, especially at 48 and 72 hours. There are a couple of reasons that explain this trend. First, a lot of research money was provided for track prediction in the 1980’s and 1990’s. This lead to the development of more robust computer model guidance, which is the main contributing factor to the increase in track skill. Another explanation is advances in understanding of what causes tropical cyclones to move in certain directions. This is partly the result of the increased resources for track prediction. A third explanation is personnel continuity at the National Hurricane Center – over time a forecaster will acquire a certain amount of intuition about TC forecasting. This will lead to better forecasts, especially for those situations where the model guidance is conflicting or ambiguous.

The trend for intensity prediction (not shown) is flat since 1970 at all forecast hours. This is rather disturbing. The deficiency in intensity forecasts stems from two things. First, our understanding of what causes intensity change is much less advanced than TC movement. If we don’t know what to look for or what to model, we cannot model it. Second, it is believed that intensity change, especially rapid intensity change, is highly correlated with the interior structure of the TC. These are areas where data is just not available in significant enough quantities to make a consistently skillful intensity forecast. Although forecasters are well aware of the importance of things like oceanic heat content, warm core eddies, concentric eyewall cycles, vertical wind shear and more, it is the complex interactions between those factors that make intensity prediction a challenging problem.
Dr. William Gray (Colorado State University)

Dr. Gray pioneered work in the prediction of seasonal tropical cyclone activity in the Atlantic with a series of papers in the 1980s and 1990s (Gray 1984a, 1984b, Gray et al. 1992, 1993, 1994). He found that there were winter and springtime signals important to tropical cyclogenesis in the climate system that would persist into the season. Those signals and physical significance for tropical cyclogenesis are summarized below:

1) **Quasi-Biennial Oscillation (QBO)** phase - The QBO is a stratospheric oscillatory wind (around 10-12 miles altitude) that reverses phase about every 13 months. When the QBO winds are in the east phase, tropical cyclone activity in the Atlantic tends to be suppressed. Gray hypothesizes that convection in the off-equatorial latitudes (8°-18°N), where genesis is common, is minimal compared to equatorial (0°-7°N) convection when the QBO is in the east phase.

2) West African Rainfall – Intense rainfall in the African Sahel in the months preceding the hurricane season tended to create stronger African easterly waves (seedlings for tropical cyclones) and were correlated with easterly wind anomalies over the Caribbean Sea (which decreased the vertical wind shear in that area).

3) Caribbean Sea Level Pressure Anomaly (SLPA) – This parameter is related to the location of the Intertropical Convergence Zone (ITCZ). If the anomaly was high (higher pressure than normal), tropical cyclone activity tended to be suppressed since the low-level convergence and vorticity associated with the ITCZ was displaced to the south.

4) Caribbean 200 mb Zonal Wind Anomaly (ZWA) – If positive, this meant that the upper-level winds were more westerly, increasing vertical wind shear over the region and thus suppressing tropical cyclogenesis. Was a good predictor because the springtime ZWA tended to persist into the hurricane season.

5) **El Niño Southern Oscillation (ENSO)** – The warm phase of ENSO (El Niño) enhances the upper tropospheric westerlies over the Atlantic Basin, limiting tropical cyclone formation.

Using these predictors, objective measures of tropical cyclone activity were computed in the December (Gray et al. 1992) preceding the season, the first day of the season (1 June, Gray et al. 1994), and before the prime development months (1 August, Gray et al. 1993). Gray’s method had shown considerable skill up until the 1995 season. That season marked the beginning of a remarkable increase in Atlantic tropical cyclone activity, capped by the record-setting 2005 season. Gray has consistently under-predicted Atlantic activity since 1995. Because of that, he and his research group revised their methodology and predictors for seasonal prediction. The 1 June predictors are based on ENSO, sea level pressure anomalies off the coast of Africa, and the phase of the North Atlantic and Arctic Oscillations. Using the new methodology, on 1 June Gray forecasted 15 named storms for the 2005 season – 60% too low.
NOAA began to issue hurricane seasonal forecasts in 1999. They base their forecasts on two predictors employed by Gray - ENSO phase and SST conditions in the Atlantic. However, they also strongly consider the phase of the Atlantic multi-decadal oscillation (AMO). The AMO is a set of atmospheric conditions over the North Atlantic that tend to occur together. Figure 17 shows the important components of the AMO. Briefly, during an active AMO, conditions are more favorable for tropical cyclone formation in the Atlantic. SSTs are warmer, vertical wind shear is less, and upper-atmospheric conditions are more favorable. It is theorized that the AMO is driven by decadal (20-30 year) changes in the Atlantic ocean circulation. Figure 18 shows marked changes in tropical cyclone frequency (here, shown in terms of what is called the “Accumulated Cyclone Energy (ACE) Index”, which is just a measure of the length of duration and strength of tropical cyclones during a given year) that are attributed to changes in the AMO. NOAA also grossly underpredicted the 2005 Atlantic hurricane season. More information on the NOAA seasonal forecast can be found at: http://www.cpc.noaa.gov/products/outlooks/hurricane.shtml.

**Review Terms**

- GOES
- Statistical Models
- Dynamical Models
- Track Models
- Nested model
- Consensus model
- RAMMB Statistical Model
Figure 17. Conditions present in the Atlantic Basin during an active multi-decadal signal. From http://www.cpc.noaa.gov/products/outlooks/figure2.gif.

Hennon Statistical Model

QBO

AMO

References


Chapter 7
Tropical Cyclone Impacts

It should come as no surprise that tropical cyclones are one of the most destructive natural disasters that humans have to confront. Although most think of the strong winds as the biggest threat, the Katrina landfall in New Orleans in 2005 should remind us that water is the biggest killer, and usually the biggest cause of damage in a landfalling tropical cyclone. Historically, fresh water flooding has killed the most people in hurricanes. This section will summarize the four deadly impacts of tropical cyclone landfalls: wind, storm surge, tornadoes, and rainfall.

Wind

Tropical cyclones create horizontal wind (the so-called “primary circulation”) through the horizontal pressure gradient created by the rising cumulonimbus clouds near the center. There is a net mass loss at the center of the TC, thus creating a relatively low pressure center there.

As we saw in the discussion of TC lifecycle, tropical cyclones are categorized by the “sustained maximum wind speed” of the system. The Saffir-Simpson (SS) scale is used to classify the intensity of the tropical cyclone. The SS scale was originally devised by Herb Saffir, a wind engineer who was studying the effects of wind damage on homes in south Florida. Realizing that there was no convenient scale to rank hurricane severity, he devised the simple “1-5” category based on damage caused by different hurricane winds. Bob Simpson, director of the NHC at the time (1969), adopted Saffir’s scale for use at NHC, adding the effects of storm surge and flooding and basing it primarily on wind speed. Figure 19 shows the categories of the SS scale and the expected storm surge by category. Storm surge is the advance of sea water over land forced by the hurricane winds, and will be addressed later in this section.

<table>
<thead>
<tr>
<th>Storm Cat.</th>
<th>Trop. Depression</th>
<th>Trop. Storm</th>
<th>Cat. 1</th>
<th>Cat. 2</th>
<th>Cat. 3</th>
<th>Cat. 4</th>
<th>Cat. 5</th>
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</thead>
<tbody>
<tr>
<td>Air pressure (mb)</td>
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<td>&gt; 980</td>
<td>965 – 980</td>
<td>945 – 965</td>
<td>920 – 945</td>
<td>&lt; 920</td>
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<tr>
<td>Wind speed (mph)</td>
<td>&lt; 39</td>
<td>39-73</td>
<td>74-95</td>
<td>96-110</td>
<td>111-130</td>
<td>131-155</td>
<td>&gt; 155</td>
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<tr>
<td>Storm Surge (ft.)</td>
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<td>--</td>
<td>4-5</td>
<td>6-8</td>
<td>12</td>
<td>13-18</td>
<td>&gt; 18</td>
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</tbody>
</table>

Figure 19. The Saffir-Simpson scale for categorizing tropical cyclones.
The types of wind damage by category are:

*Category 1:* Unanchored mobile homes, shrubbery, small trees, coastal flooding
*Category 2:* Roofing, door, window damage, considerable tree damage
*Category 3:* Structural damage to small buildings, trees blown down or foliage lost, mobile homes destroyed
*Category 4:* Total roof failures, all signs/trees blown down, extensive damage to all buildings
*Category 5:* Complete roof failures on many buildings, some buildings completely destroyed, severe and extensive window/door damage to other buildings.

It should be noted that the amount and types of damages to structures are highly dependent on the quality of construction and the implementation and enforcement of building codes. Figure 20 shows some examples of wind damage from each category.

*Figure 20: Examples of TC wind damage.*
Storm Surge

Storm surge is a rise in sea level attributed to the “piling up” of water by the winds of the tropical cyclone and, to a much smaller extent, a raised dome of water from the extreme low pressures under the TC center. The definition of storm surge does not include the waves on top of the sea level rise, which in themselves can exceed 30 ft. in height. Combined with high tides, storm surges can cause extreme devastation in coastal areas.

Typical storm surges are several meters in depth, although the strongest hurricanes can produce storm surges in excess of 10 meters (~30 ft.). The amount of storm surge experienced by a coastal area depends upon several factors: strength and extent of the wind field, translation speed of the storm, coastline shape, and the slope of the continental shelf. In general, for a given storm, areas located to the right of the landfalling hurricane where the coastline forms an inverted “U” shape will tend to experience the highest surges.

Examples of Devastating Storm Surges

The largest loss of life (at least known) in the United States from storm surge was the 1900 Galveston hurricane. Thought to be a Category 4 storm that came ashore with little warning, this hurricane threw 30 ft. of the Gulf of Mexico over the city, claiming an estimated 6,000 lives. Figure 21 is a photo taken after the passage of the hurricane.

Another large storm surge event occurred with the landfall of Hurricane Camille (1969). Camille is one of only 3 Category 5 hurricanes known to have made landfall in the United States. The storm surge from Camille was estimated to be more than 7 meters above sea level (more than 2 stories), with large breaking waves on top. 100 deaths and $1 billion damage was attributed to Camille’s surge alone. Figure 22 shows famous photos of the Richilieu complex in Pass Christian, MS. Of the two dozen occupants at the time of landfall, only 1 survived – found in a tree 5 miles inland.

Globally, the largest loss of life from storm surge occurred in November 1970 in Bangladesh. An estimated 9 meter (~27 ft.) surge at Chittagong Bangladesh claimed an estimated 350,000 lives. Another cyclone 21 years later killed another 140,000 people. Storm surges in that region of the world are enhanced by the confluence of coastline and the very low elevation (only a few feet above sea level) of the country.

SLOSH – Modeling Storm Surges

Forecasters have a very good tool at hand for predicting how high the storm surges for a particular storm in a particular area will be. The SLOSH (Sea, Lake and Overland Surges from Hurricanes) model is run at the NHC and estimates storm surge based on the following variables: Storm pressure, size, forward speed, track, wind speed and extent, and coastline features (shape, bay and river configurations, water depths, bridges, etc.).
Figure 21. Galveston, TX after the 1900 hurricane. Damage shown in this photo was caused by storm surge.

Figure 22. The Richelieu Apartments in Pass Christian MS before Camille (left) and after Camille (right). 23 of 24 occupants were killed in the 7 meter storm surge.

The SLOSH predictions are accurate to within 20% of actual surges. The accuracy is highly dependant on the accuracy of the forecast track. One can imagine how a shift of a few tens of miles may place a location on the right hand side rather than left hand side of the storm, drastically changing the potential for severe storm surge. Because of this, the SLOSH model is
best used to warn of the maximum potential surge for a location. This information is commonly used by local emergency management officials in determining who and when to evacuate.

**Tornadoes**

A study was published in 1983 (Gentry 1983) that documented the occurrence of tornadoes in landfalling hurricanes between 1960-1982. His findings were:

- Nearly every hurricane had tornadoes associated with it during landfall
- 60% of tropical storms produced tornadoes
- Virtually all tornadoes occurred in the right-front quadrant of the storm
- Most are weak, but some tornadoes achieved F-3 strength
- 80% of tornadoes formed in the outer rainbands, 20% near the outer eyewall

Figure 23 shows the area of tornado occurrence relative to the storm motion. As Gentry notes, the front right quadrant is by far the most common area for tornadogenesis. It is thought that this area of the storm creates the most shear in the horizontal wind from the surface (which is drastically slowed by the land area) to the upper storm levels (not yet affected by the land). This sets up a horizontally oriented circulation between the levels, as shown in Figure 24. If a strong updraft in the rainband pushes the circulation vertically, it can result in a tornado.

![Figure 23. Display of locations of 1973-1980 tornadoes along the direction the tropical cyclone was moving. The dashed curve encircles 95% of the observation points (From Gentry 1983, Copyright American Meteorological Society).](image)
Figure 24. A schematic drawing that shows how boundary layer friction can induce circulation in a landfalling hurricane. As the storm approaches the shore (a), the ocean surface provides relatively little frictional resistance and thus there is little vertical wind shear. As the circulation moves over land (b), land features create a large drag on near-surface winds as the upper-level winds still remain vigorous. This speed shear induces a horizontally oriented circulation.

There is a preferred time of day for tornado formation. They form most frequently between the hours of noon and 6 pm (local time). This is shown in Table 2 below. It is thought that there is enough daytime heating before the noon hour to destabilize the atmosphere enough to create more upward motion. This makes the tilting of the shear-induced circulation described above more likely.

A previous study reported that tornadoes contribute up to 10% of all fatalities in a landfalling hurricane and up to 0.5% of the total damage. So tornadoes certainly are not a trivial consideration in TC landfalls, especially in tropical storms where they may cause proportionally more damage and fatalities.

<table>
<thead>
<tr>
<th>Time (LST)</th>
<th>Percentage of tornadoes</th>
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<tr>
<td>00-03</td>
<td>6</td>
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<tr>
<td>03-06</td>
<td>9</td>
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<td>06-09</td>
<td>4</td>
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Table 2. Tornadoes associated with hurricanes and time of day (From Gentry 1983, Copyright American Meteorological Society).

**Rainfall**

Tropical cyclones can dump a copious amount of rainfall, leading to the potential for a large loss of human life. Hurricane Mitch (1998), already setting records as the latest Category-5 hurricane in the Atlantic season, moved slowly ashore into Central America and killed 11,000 people over a few days from mudslides caused by extreme rainfall.

Cerveny and Newman (2000) undertook a study of tropical cyclone rainfall as derived from microwave satellites. They examined 877 tropical cyclones from 1979-1995. They found a high correlation between the average rainfall per day (mm/day) and the maximum wind speed of the
tropical cyclone. Figure 25 shows this result. Note that category-4 and 5 hurricanes produce on average 350 mm (35 cm, or over 1 ft.) of rainfall in a single day.

Figures 26 and 27 also yield interesting information about tropical cyclone rainfall. Fig. 26 shows the percentage distribution of rainfall from near the center of the system. Overall, the inner core of the TC produces about 25% of the total rainfall of the storm. However, that percentage is higher both for weaker systems (> 30%) and category-5 storms (~35%). Fig 27 shows that systems located deeper in the tropics create on average rain rates that are twice that (~225 mm/day vs. 100 mm/day) of storms in the sub-tropics or mid-latitudes. This is probably due to the weakening generally experienced by storms as they move over cooler waters.

The biggest threat of flooding to inland areas with everything else being equal is the translation speed of the storm. Slow moving systems, even those of tropical storm or depression strength, have been observed to drop more than 30” of rainfall over localized areas. Tropical Storm Allison (2001) dropped up to 37” of rain in the Greater Houston area, causing massive flooding (Figures 28 and 29).

---

![Graph](image.png)

*Figure 25. Relationship between the average tropical cyclonic rainfall (mm day$^{-1}$) over nine 2.5° x 2.5° grid cells and the maximum surface wind speed (in 10-kt. categories) for tropical cyclonic observations of the Atlantic and North Pacific basins (From Cerveny and Newman 2000, Copyright American Meteorological Society).*
Figure 26. Ratio of the center grid cell rainfall to the average tropical cyclonic rainfall (in percent) stratified by maximum surface wind speed (10-kt categories). Bars indicate one standard error deviation about the mean.

Figure 27. Latitudinal gradient in average tropical cyclonic rainfall (mm day$^{-1}$) associated with the North Pacific and Atlantic basins. Bars indicate one standard error deviation about the mean.

Figure 28. Downtown Houston TX after Tropical Storm Allison (2001). From hydrology.rice.edu/flood/photos/index1.htm.
**Study Terms**

Saffir-Simpson Scale

Storm surge

SLOSH

**References**


Observing the Tropics

Introduction

As we have already seen in regards to hurricanes, observation of the tropical atmosphere is a challenge. Much of the action is over the world’s oceans, where surface observations are non-existent beyond a few ship reports and buoy observations here and there. Thus we rely heavily on remotely sensed data, both of the atmosphere and ocean.

This section will highlight five (5) methods of obtaining data in the tropics beyond surface and upper air stations. Please note that there are many more sources of observations, but those that follow are most heavily relied upon for observation and prediction of tropical cyclones and as input into global forecast models.

Buys

Much of the tropics are water, and thus it is difficult to obtain a set of high-density surface observations. Beyond a few ship reports and data from satellites (of which a few are discussed later in this section), the only data obtained from the vast oceans is from buoys.

Types of Buoys

A drifter is a buoy that is attached to a sub-surface sea anchor (called a “drogue”). It floats freely on the water surface, transmitting observations of temperature, humidity, salinity, wave heights, etc. to passing satellites. Each of those satellites contain a system called Argos, which collects the data, determines the identity and location of the buoy, and then sends the data to the Atlantic Oceanographic and Meteorological Laboratory (AOML) for further quality control and processing. More information can be found at http://www.aoml.noaa.gov/phod/dac/gdp_drifter.html or Lumpkin and Pazos (2006). See http://www.aoml.noaa.gov/phod/graphics/dacdata/globpop.gif for a map of current locations of global drifters.

A moored buoy is one that is anchored to the sea floor and is thus not moving with the wind and ocean currents. Moored buoys serve an important purpose beyond routine monitoring of local conditions. They provide vital information on approaching tropical cyclones, including wind speeds, wave heights, and pressures.

TAO Array

The National Data Buoy Center (NDBC) collects, checks, and distributes moored buoy data from around the world. Figure 29 shows the regions of moored buoys under NDBC control. One of the more critical “arrays” of buoys is the TAO array in the Pacific Basin. The TAO array
plays an important part in the detection and observation of the El Niño – Southern Oscillation (ENSO, discussed later). It consists of 70 moorings in the tropical

Figure 29. Available moored buoy data available at the National Data Buoy Center (from http://www.ndbc.noaa.gov/rmd.shtml).

Figure 30. Observations of sea surface temperature and wind from the TAO array ending 30-May 2006. The top plot are the observations, the bottom is the departure from normal. During this time much of the Pacific was in a neutral ENSO phase.
Pacific that take sea surface temperature observations at regular intervals. Figure 30 shows the SST conditions and winds at the end of May in 2006. Note that the easterly trades were slightly weaker than normal at this time, while the SSTs were near average.

**Reconnaissance Flights (Tropical Cyclones)**

Since 1944, reconnaissance missions have been flown into tropical cyclones. These flights are critical for the operational analysis and forecasting of tropical cyclones. Aircraft take in situ measurements of flight level winds, pressure, and temperature. More importantly, they routinely drop GPS Dropsondes (Figure 31). These devices fall to the ocean surface, measuring a number of meteorological variables as they drop. These data are transmitted in real-time to a receiving station at the National Hurricane Center, where they can be made available to forecasters just minutes after they were taken from inside the tropical cyclone.

Reconnaissance missions are flown by the 53rd Weather Reconnaissance Squadron (United States Air Force), known as the “Hurricane Hunters”. They are based at Keesler Air Force Base in Biloxi, Mississippi. The squadron consists of 10 specially configured WC-130 aircraft (see Figure 32), and is also tasked to fly wintertime storms off the northeastern U.S. coast.

**Tropical Rainfall Measuring Mission (TRMM)**

TRMM is the first space-borne precipitation radar that measures the vertical distribution of precipitation over the tropics (between about 35°N and 35°S). It was launched on November 27, 1997. Up until that time, global estimates of tropical precipitation were in error by as much as 50% for reasons including limited in situ observations and errors in those observations being made due to the convective nature of the rainfall. TRMM has drastically reduced those errors. This is important for a number of reasons; the first is improving estimates of the Earth’s energy budget, of which a large portion of heating is caused by latent heat release in rainfall. Other applications of TRMM data include obtaining surface soil moisture data (Gao et al. 2006), lightning research (Petersen et al. 2005), detection of the Madden-Julian Oscillation (Cho et al. 2004, see Section III), and improvement of global sea surface temperature (SST) datasets (Reynolds et al. 2004).

But perhaps the most visible use of TRMM products has been in the application to tropical cyclones. One benefit to the TRMM satellite is that it contains both active and passive sensors onboard. “Active” means that the sensor sends out a pulse, collects the return signal, and then processes the information. “Passive” means that the sensor does not actively send out a signal, but rather measures radiation emitted naturally from the earth/atmosphere system.
Figure 31. Schematic diagram of the GPS dropsonde (http://www.gsfc.nasa.gov/gsfc/earth/pictures/camex4/dropsonde.gif).

Figure 32. A WC-130 Hurricane Hunter aircraft (http://www.hurrican hunters.com/phstart.jpg).
The active sensor aboard TRMM is the precipitation radar (PR), which provides accurate information about rain rate. The passive sensor is called the TRMM Microwave Imager (TMI). The TMI provides information that is used to fix the center of tropical cyclones. Figure 33 shows a TRMM pass over Tropical Cyclone Manou, which is approaching the Madagascar coast. The TMI has a much wider swath than the PR. Note the precipitation echoes in the inner PR swath. The yellow areas in the outer swath indicate precipitating ice particles in rainbands away from the center of Manou.

Figure 34 shows an example of how the TMI can identify centers of circulation in tropical cyclones, even when one may be hidden by a cirrus canopy or convective overshoots. This image is of the Brazilian hurricane that made landfall in March of 2004.

Although the TRMM satellite completed its full mission life in 2004, supplemental funding has been awarded to keep it in orbit until its fuel runs out. It is hoped that the next generation of precipitation satellites (Global Precipitation Mission, or GPM) is in place by that time so that the benefits of having such a system continue.

**QuikSCAT**

Another polar orbiting satellite that has become very useful in recent years is the SeaWinds scatterometer aboard the QuikSCAT satellite. A scatterometer is an active instrument that sends out a pulse of radiation toward the ocean surface. The pulse is backscattered by capillary waves (very small ripples on the water surface) back toward the satellite. The amount of backscatter is directly correlated to the near-surface wind speed – more roughness is caused by higher wind speeds, which in turn will increase the backscatter. It is also possible to determine the near-surface wind direction from QuikSCAT. This is determined by analyzing the direction in which the capillary waves are oriented. Typically, QuikSCAT will measure the backscatter from four different “look angles”. As long as uncontaminated (e.g. no rain) measurements can be determined from each of these angles, it is possible to get an accurate wind direction.

*Sensitivity to Rain*

There are some limitations to QuikSCAT – mainly, its accuracy is severely degraded in intense rain. Rain disrupts the “true” backscatter in three ways:

1. Enhances backscatter by downward oriented QuikSCAT pulse
2. Attenuates (absorbs) radiation backscattered from the ocean surface
3. Raindrop “splash” on the surface will artificially enhance backscatter

These responses compete with each other and it is generally not known how large or small the net impact will be. In general, when the true surface wind is light, rain will make the satellite interpret the wind to be much stronger than it actually is. Conversely, a strong true wind will generally be in error on the low side, as attenuation tends to dominate in this situation. Rain also negatively impacts the determination of a wind direction, as the character of the surface backscatter becomes unclear.
Figure 33. TRMM pass over Tropical Cyclone Manou. The inner swath (PR) shows rainfall rates (red = highest); the outer swath (TMI) detects rainbands, and can also be used to detect circulation centers.

Figure 34. TMI pass near landfall of the 2004 Brazilian hurricane. Red areas represent strong convection. The center of the circulation is noted with an arrow.
How QuikSCAT is Used

Figure 35 shows a QuikSCAT swath over the western Atlantic Ocean during March of 2004. The end-product is a series of wind barbs. Blue colors are generally light winds,
Figure 36. QuikSCAT pass on 8-September 2003 over TD-14 in the eastern Atlantic. Note how QuikSCAT identifies the circulation center that is invisible in the Infrared satellite image.

**Advanced Microwave Sounding Unit (AMSU)**

**AMSU** is a passive radiometer that flies onboard a suite of polar orbiting NOAA satellites (NOAA-15, NOAA-16, and NOAA-17). It senses emitted radiation in the microwave spectrum (wavelengths on the order of centimeters). The uniqueness of this instrument is that it can determine temperatures of many layers of the atmosphere (hence the name “Sounding Unit”) through the use of 15 different channel sensors.

Four of those channels (channels 5-8) are used to detect the existence of a warm core in tropical cyclones. This is a critical observation that is used to decide whether a suspicious system has the characteristics of a tropical cyclone or a cold-core system such as a mid-latitude cyclone. Figure 37 shows a cross section of temperature in Hurricane Floyd (1999) from an AMSU pass. Note the 10°C+ warming anomaly indicated over the center of Floyd at around 300 mb.

AMSU is also used to estimate the intensity of a tropical cyclone. Researchers at the University of Wisconsin have found that AMSU channels 7 and 8 are highly correlated with TC intensity. Table 3 below shows that the UW algorithm outperformed the traditional *Dvorak technique* in both the Atlantic and Northwest Pacific basins during the 2004 season (see [http://amsu.ssec.wisc.edu/explanation.html](http://amsu.ssec.wisc.edu/explanation.html) for more information).

<table>
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<th>Dvorak</th>
<th>AMSU</th>
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Table 3. Comparison of AMSU and Dvorak intensity errors for the 2004 season. Data were validated against reconnaissance measurements within 3 hours of the pass time in the Atlantic. In the NW Pacific, data were validated against drifting buoys, ship observations, and surface observations. From http://amsu.ssec.wisc.edu/explanation.html.

<table>
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<tr>
<td>RMSE</td>
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<td>4.7 mb</td>
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Figure 37. AMSU derived temperature anomaly (K) for Hurricane Floyd (1999). From http://amsu.ssec.wisc.edu/explanation.html.
Study Terms

buoys

drifter

Argos

moored buoy

NDBC

TAO Array

GPS Dropsondes

TRMM

Active vs. Passive

Precipitation Radar (PR)

TRMM Microwave Imager (TMI)

Scatterometer

QuikSCAT

backscatter
AMSU

Dvorak Technique

References


Chapter 9
Wind Analysis of the Tropics

Introduction

Compared to the mid-latitudes, the tropics have very weak temperature and pressure gradients. Thus, conventional analyses of isobars and isotherms are not usually possible. We usually look to wind data and analyses to locate waves, cyclones, anticyclones, etc.

Traditionally, meteorologists have analyzed wind data for direction (isogon, or streamline) and speed (isotach). We will also look at two other variables: streamfunction and velocity potential. The stream function is an indication of the rotational component of the flow, while the velocity potential identifies areas of convergence and divergence.

Types of Analyses

Isogons and Streamlines

Isogon – Line of constant wind direction
Streamline – A line along which, at any given instant, all velocity vectors at point through which the line passes are tangent to the line.

At a given instant, the flow is parallel to the streamline at a fixed point.

If we are provided a map of wind direction (either wind barbs or numerals), follow these steps to creating a streamline analysis:

1) Do an isogon analysis – contour (usually in 30° increments) the wind direction
2) For each isogon, draw tiny cross ticks at regular intervals in the direction of the wind flow. For example, if you are drawing ticks across the 270° (West wind) line, your ticks would look like this:
3) Connect the tick marks with a line – these are your streamlines.
4) Label centers of cyclones (“C”) and anti-cyclones (“A”)

There are a few things to note. Quite frequently you will see areas where your isogons come together, like the following example:

![Streamline example](image1)

This pattern represents an area where the wind speeds in the center are theoretically zero. This could be the center of a cyclone or anti-cyclone. The streamline analysis of this isogon pattern will look like:

![Streamline example](image2)

There is another possibility that the centers of isogons represent what is called a col area. This is an area that is the “saddle point” on a pressure surface. It may represent a nascent frontal zone. A col area looks like this in a streamline analysis:

![Streamline example](image3)
Of course, isogon and streamline analyses are now done primarily by computer.

Isotachs

An **isotach** is a line of constant wind speed.

Anyone who has taken analysis class should be familiar with an isotach analysis. It is straightforward – just like contouring any other variable.

Stream function

More obscure than streamlines and isotachs is the **stream function**. A stream function is a way of visualizing the wind direction. It is valid for 2-dimensional, non-divergent flow. In other words, stream functions represent only the rotational component of the wind, not the divergent part. Since 80-90% of the wind flow is rotational, especially away from the surface, stream functions account for a large percentage of the total wind.

The stream function is represented by the Greek letter “psi” (ψ). Mathematically, the stream function is defined as:

\[ \psi = \psi(x, y) \]  
\[ u = -\frac{\partial \psi}{\partial y} \]  
\[ v = \frac{\partial \psi}{\partial x} \]

where \( u \) = zonal (east-west) wind, and \( v \) = meridional (north-south) wind. So we can also see that the gradient of the stream function (\( \nabla \psi \)) provides information on the wind speed. The strength of the zonal wind depends on the gradient in the y-direction, while the strength of the meridional wind depends on the gradient in the x-direction.
The stream function \( \psi \) is constant along a streamline (for 2-D non-divergent flow), with low values of \( \psi \) to the left of the flow direction and high values of \( \psi \) to the right. Let’s look at a couple of examples.

In the figure below, we have a west wind (blowing from the west) bounded by two streamlines. A west wind is a positive zonal wind \((u > 0)\).

Since the low values of \( \psi \) are to the left of the flow, \( \psi_2 > \psi_1 \), which means that \( \frac{\partial \psi}{\partial y} < 0 \).

Therefore \( u > 0 \), since \( u = -\frac{\partial \psi}{\partial y} \) from equation (1) above. The same example can be give for a positive meridional wind, which is a south wind (flowing toward the north):

Here, \( \frac{\partial \psi}{\partial x} > 0 \) since \( \psi_2 > \psi_1 \). And by definition, we know that \( v = \frac{\partial \psi}{\partial x} \) (positive).

*Velocity Potential*

The stream function only applies to the non-divergent component of the wind flow. However, there are many interesting types of flows that contain divergence. The velocity potential is a
variable that represents the 2-D divergent component of the flow. It is represented by the Greek letter “chi” ($\chi$). Mathematically, the divergent component of the wind flow is:

$$\vec{V}_x = \nabla \chi = \frac{\partial \chi}{\partial x} + \frac{\partial \chi}{\partial y}$$

For the stream function, we saw that the total wind ($\vec{V}$) was parallel to isopleths of stream function with the low values to the left. $\vec{V}$ is perpendicular to isopleths of velocity potential and is directed from low to high values (Figure 38).

![Figure 38. Schematic plot of velocity potential isopleths (light solid) and the total wind (dark solid). Velocity potential is increasing to the right ($\chi_4 > \chi_1$).](image)

Note that the gradient of the stream function is perpendicular to the gradient of velocity potential ($\nabla \psi \perp \nabla \chi$).

**Review Terms**

- stream function
- velocity potential
- isogon
- streamline
- col area
Problems

1. What are the mks units of $\psi$?

2. Analyze the given map for streamlines. Identify centers of cyclones (“C”) and anti-cyclones (“A”).

3. Analyze the given 200 mb data for velocity potential. Shade areas of divergence in a read color and areas of convergence in blue.
Chapter 10
Climatology of the Tropics

This section will examine some of the climatological (long-term average) conditions of the tropics via the following variables: land/sea distribution, temperature and temperature gradient, atmospheric soundings, temperature range, zonal wind, cloud cover, relative humidity, and precipitable water.

Land/Sea Distribution

Figure 39 shows the percentage of land/sea in each hemisphere, divided by 5° latitude intervals. If we only examine the tropics (within 25° of the Equator), we find that the amount of land in each hemisphere is about equal. The boxes at left on Fig. 39 show the percentage of water in each hemisphere as a whole. The Northern Hemisphere has about 40% land vs. 20% for the Southern Hemisphere.

Temperature and Temperature Gradient

Figure 40 is a meridional cross section of temperature (top) and temperature gradient (bottom). A meridional cross section is created by averaging around a latitude circle (“zonal average”). Mathematically, a zonal average of ‘x’ is represented as:

\[
[x] = \frac{1}{2\pi} \int_{0}^{2\pi} x d\lambda
\]

where \(2\pi\) radians = 360°, and \(\lambda\) = longitude. A zonal average smoothes out any
variations across a latitude circle due to local and regional differences (land/sea distribution, topography, etc.). It provides us with a big picture view.

Figure 40. (Top) Zonal average temperature cross section for Northern Hemisphere winter (January through March). (Bottom) Gradient ($\nabla T$) of temperature ($\frac{\partial T}{\partial y}$) in winter.

The “gradient” of a variable represents how quickly a variable changes with distance. For the cross section in Fig. 40 above, the gradient of temperature is simply:

$$\nabla T = \frac{\partial T}{\partial y}$$

which can be thought of as “the change in temperature per distance in the meridional (y) direction. Several things can be learned from Fig. 40:

1) the tropical tropopause (approximately where temperatures approach -60°C) is much higher than the non-tropical tropopause – more on this later;
2) during winter, the warm maximum at the surface is at approximately 5°S;  
3) and, there is little temperature gradient in the tropics. The maximum gradient at the surface is around 30°N and 40°S.

The same figure for the Northern Hemisphere summer is shown in Figure 41 below.

![Figure 41. As in Figure 40 except for summer.](image)

From Figure 41 we see that:
1) the tropics have become more unstable. The upper tropospheric temperatures are colder than winter, and surface temperatures in the tropics are warmer overall;  
2) this implies stronger rising motion in the Northern Hemisphere during summer (e.g. a more robust Hadley circulation); and,  
3) temperature gradients in the Northern Hemisphere near the surface are much weaker.
Temperature Soundings

Figure 42 shows mean temperature soundings by latitude. The 0° and 20° soundings can be considered to be tropical soundings, while the others are extratropical. The heights of the tropical and polar tropopause are noted. The tropopause is usually defined by the level at which the temperature profile becomes isothermal. In the tropics, this generally occurs around 100 mb (16 km) – in the extratropics, the tropopause averages about 300 mb (10 km) in height. This is a direct consequence of the average temperature in the atmospheric layers. The hypsometric equation tells us that the thickness of the layer is directly proportional to the mean virtual temperature. As seen from Figure 42, a warmer mean temperature produces a thicker

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Figure 42. Mean temperature soundings for Northern Hemisphere summer (June – August) for a number of latitudes.
layer. The above Northern Hemispheric soundings are for the Northern Hemisphere summer – how would the soundings change for wintertime?

Figure 43. Meridional cross section of the zonal wind (m/s) for Northern Hemisphere winter (top) and summer (bottom). Shaded areas (negative values) represent easterly winds. Isotherms are dotted (From Holton 2004).

Zonal Wind

Figure 43 is a meridional cross section of the zonal wind (west-east wind). During the Northern Hemisphere winter (top), easterlies on average extend to about 30° latitude near the surface. The NH polar jet stream can be seen at about the tropopause level (300 mb) and is about twice as strong in the winter as summer (42 m/s vs. 21 m/s).

When the zonal wind is shown on a plan map (Figure 44), one can see the upper tropospheric westerlies are quite strong in the NH during winter (top panel). This creates severe vertical wind shear, as the surface winds (not shown) are still generally easterly in the tropics. Thus, tropical cyclone formation is prohibited during this season although SSTs are still marginally favorable
over a broad area in the western Pacific and even Atlantic basins. In rare cases, tropical development has occurred during the NH spring.

Figure 44. Mean Northern Hemisphere winter (top) and summer (bottom) 200 mb zonal wind (m/s). ‘W’ labels westerly wind maxima, ‘E’ highlights easterly wind maxima. Easterly winds are dotted contoured, westerly winds are solid.

months when a low became cut-off from the flow and acquired tropical characteristics (e.g. Tropical Storm Anna (2003) developed in April). During the NH summer, the 200 mb westerlies are much weaker or even easterly over the prime TC development regions of the western Pacific and Atlantic. In fact, easterlies are prevalent at every longitude within about 10° of the Equator. This flow is created from the turning of the ITCZ outflow to the right by the Coriolis force – the ITCZ is located at an average latitude of 10°N during the summer months.
Figure 45. Mean annual temperature range (°F).

Annual Temperature Range

Figure 45 above shows the average temperature range for the globe. ‘X’s mark areas of maximum temperature range, and ‘O’s mark regions of minimum temperature range.
Temperature range is simply the difference between annual maximum and minimum temperatures. Note that the tropics generally have very small temperature ranges, and that the areal extent of the low ranges is much higher over the oceans (which change temperatures rather slowly due to a low specific heat, among other things). By far the highest temperature ranges occur over the interior of extratropical continents (e.g. North America, Africa, and Asia). Maxima in the Saharan Desert are due to intense radiational imbalances of the sandy surface between seasons and diurnally.

Cloud Cover

Figure 46 shows the total cloud cover (%) for winter (top) and summer (bottom). During NH winter, the maximum (filled with dots) seen in the North Pacific, Atlantic, and over the entire extratropical SH are due to stratus cover. Cloud cover minima (stipled) are found over the Saharan Desert and south Asia. For the NH summer, large minima areas
are seen in the climatological locations of the sub-tropical highs, namely the Middle-East, Australia, and south Africa. The wet phase of the Asian monsoon circulation (more on that later) gives rise to the maximum in cloud cover seen in south Asia from eastern India up into China.

**Relative Humidity**

**Relative humidity** indicates how close the atmosphere is to saturation. Recall that it depends on air temperature as well as the specific humidity of the atmosphere. Figure 47 shows the NH wintertime relative humidity (RH) for much of the troposphere. In the tropics, the low levels are moist, particularly in the rising branch of the ITCZ which is centered around 5°S at this time. Two areas of dryness appear at around the 500 mb level. These correspond to the sinking branches of the tropical Hadley cells. Of course, sinking air warms adiabatically, which will lower the RH assuming the moisture content of the air remains constant. During the NH summer, the center of action shifts northward.
Precipitable Water

**Precipitable water** (PW) is the vertically integrated specific humidity in an atmospheric column. If all of the moisture were rained out of the column, it would have a depth equivalent to the values in Figure 48 (in units of mm). Figure 48 shows that the PW maximum corresponds to the axis of the ITCZ (shown by the dotted lines). This is the area of low-level surface convergence, whereby the trade winds advect moisture into the convergent zone. Note that there is a fairly symmetrical decrease of PW as one moves away from the tropics, with variations due mostly to topography considerations (e.g. Rockies in North America, Himalayans in Asia).
Figure 49 below shows the monthly mean precipitation for January (top) and June (bottom) 2003. This data is from the TRMM satellite, discussed in section II.

![Map showing precipitation patterns with annotations for ITCZ, SPCZ, and Asian monsoon flow.]

A few patterns of the tropical rainfall will be noted here and discussed in more detail later on. The ITCZ is clearly visible in the precipitation pattern for both months. The location of the ITCZ does not move appreciably in the Pacific basin between January and June but is noticeably farther south in January in the Indian and Atlantic Oceans and (even more so) over the continents. The South Pacific Convergence Zone (SPCZ) also shows up well in the mean precipitation maps, with more robust rain rates in the SH summer season. Finally, the Asian monsoon is clearly seen. During the NH winter, there is virtually no precipitation over south Asia. After the changeover in spring, south Asia experiences the highest mean rain rates than any other location in the world, with rates exceeding 20 mm/day on average.

**Climate Change and the Tropics**

It is now almost universally accepted that the Earth’s climate is currently changing very rapidly and that it is caused by the addition of greenhouse gases (carbon dioxide, methane, CFCs, etc.) by humans. The latest report from the Intergovernmental Panel on Climate Change (IPCC – see [http://www.ipcc.ch](http://www.ipcc.ch)) states that “warming of the climate system is unequivocal, as is now evident
from observations of increases in global average air and ocean temperatures, widespread melting of snow and ice, and rising global average sea level”. The IPCC is a collection of hundreds of climate scientists from around the world who work towards a consensus on the state of the climate and the impacts of climate change. All projections presented in this chapter originate from the 2007 IPCC synthesis report.

The first critical point to note is that “global warming” does not mean that every location around the globe has shown or is predicted to show a warming trend. There are several locations (mostly in the southern hemisphere) that have cooled in recent decades. And when one compares the warming trends of the tropics vs. the polar latitudes, the tropics have warmed much more slowly. This is due primarily to the ice-albedo feedback mechanism in the polar regions; warming melts the high-albedo ice, which in turn lowers the ground albedo, which thus allows for more solar absorption and more warming, which then melts more ice, and so on.

Nevertheless, the tropics have been warming in recent decades, especially over the continental regions. The remainder of this section will try to answer two questions:

1. How are temperature and precipitation trends in the tropics going to change? 2. How will tropical cyclone frequency and intensity change?

*Tropical temperature and precipitation changes*

In order to predict climate change, it is necessary to define different scenarios of greenhouse gas emissions. For example, will humans continue with “business as usual” emissions? Or will there be drastic cuts in emissions? Will emissions accelerate? Climate change predictions are made for several different scenarios in order to produce some sort of bounds on the predictions.

The left panel of Figure ? shows 4 scenarios for future greenhouse gas emissions and the projected global temperature increase. Regional changes for 3 scenarios (A2, A1B, and B1 – see the report at http://www.ipcc.ch/pdf/assessment-report/ar4/syr/ar4_syr.pdf for more information) are shown at right. Note that the Northern Hemisphere polar regions are projected to have the most warming (6°-7° C!), while the tropics as a whole are projected to warm around 2°C (with continental areas warming more). Figure ? shows model predictions of changes in precipitation for the period of 2090-2099 compared to 1980-1999. Note that the ITCZ appears more vigorous, with precipitation increases of 10-20%. In the subtropics, we see a marked decrease in precipitation. Also note the large increases in the polar regions, due to higher moisture capacities of the warming air.
Changes in Tropical Cyclones in a Warming World

In the wake of the Hurricane Katrina disaster, two papers were published that rocked the tropical cyclone community. Webster et al. (2005) used the “best-track” database and counted the numbers of storms over the last century. They concluded that there is a global “30-year trend toward more frequent and intense hurricanes” that “is not inconsistent with recent climate model simulations that a doubling of CO₂ may increase the frequency of the most intense cyclones”. Figure ? (at left), from their paper, shows a large increase in the number of category 4 and 5 storms since 1970.

The second paper was written by Kerry Emanuel, a well-respected tropical scientist (Emanuel 2005). He examined changes in Power Dissipation of tropical cyclones from about 1930 – 2005. Power dissipation (PD) is defined as:
where the maximum sustained wind $V$ (from the best track database) is cubed and summed over the lifetime of the storm. The wind is cubed because it has been shown that monetary losses from TCs increase roughly with the cube of the wind speed. The PD has units of energy, and can be thought of as the total power that the storm dissipates over its lifetime.

Emanuel shows that the PD has increased significantly for the North Atlantic and Pacific basins since around 1970 (Figure ?). Furthermore, there appears to be a high correlation between PD and SST for each basin. Emanuel concludes that “the near doubling of power dissipation over the period of record should be a matter of some concern”.

So what’s the controversy? SSTs have been increasing since 1970, so shouldn’t TCs naturally increase in intensity and frequency? Well, it isn’t quite that simple. First, TCs are controlled by a bunch of factors beyond the SST. Intensity wise, it has been theoretically shown by Emanuel that the difference in the upper tropospheric temperature and SST is the critical consideration for TC intensity, not just the SST. Furthermore, one has to consider other factors such as vertical wind shear, dynamical constraints (e.g. concentric eyewall cycles), atmospheric moisture (SAL), and so on. In fact, recent modeling studies by Emanuel and others (Emanuel et al. 2008, Knutson et al. 2008) have shown that TC frequency will decrease over the next century, due primarily to increases in vertical wind shear in development regions. Knutson et al. (2008) also show that hurricanes should increase in intensity overall (with some regional differences), show increased rain rates, but would not expand their development areas.

Another fundamental counterargument to the 2005 Webster and Emanuel papers is that both used “best-track” databases for their analyses. The best track databases are flawed, especially prior to the satellite era (mid-1960s). Many storms, especially those that did not approach land, were undoubtedly missed in the first half of the 20th century. Intensities of storms that do
appear in the database prior to the 1970s were most likely underestimated due to under (or no) sampling of the storms. Even after the mid-1970s and up to the present day, most storm intensities are estimated via the Dvorak technique, which provides an estimate of the maximum wind based on a single IR satellite pattern. Although the Dvorak technique has been shown to provide decent estimates of maximum wind, it has flaws, especially in cases where there is rapid intensification or dissipation of the TC.

Summary

This section introduced you to some of the climatological features of the tropics as well as globally, including how these features may change in a changing climate. The last third of the course will be dedicated to examining the phenomena that are responsible for creating these patterns.

Review Terms

Temperature range

Relative humidity

Precipitable water

Power Dissipation

Dvorak Technique

References


reduction in Atlantic hurricane frequency under twenty-first-century warming conditions. [http://www.nature.com/naturegeoscience](http://www.nature.com/naturegeoscience).

Chapter 11
Tropical Field Experiments

One of the most exciting things that one can do in science is to be a part of a field experiment. This involves going out into the field (the tropics!), taking in depth measurements, collecting the data, saving it, and analyzing it at a later date. In recent decades, there have been numerous field experiments in the tropics that have yielded a large amount of data and new understanding. This section will highlight three such experiments: GATE, TOGA-COARE, and TCSP.

**GARP Atlantic Tropical Experiment (GATE)**

One of the most fun aspects of a field experiment is to come up with an acronym that is cool and actually says something about what you are trying to accomplish. GATE is a nested acronym—the full acronym is “Global Atmospheric Research Program Atlantic Tropical Experiment”.

The purpose of GATE was “to understand the tropical atmosphere and its role in the global circulation of the atmosphere” (http://www.ametsoc.org/sloan/gate/index.html). It was a huge undertaking. It involved 40 research ships, 12 research aircraft, many buoys, and involved 20 different countries. The experiment took place during the summer of 1974 and covered the tropical Atlantic Ocean from Africa to South America. It is estimated that over 1000 scientific papers have been published using GATE data – research continues today.

*Important Findings of GATE*

One feature of the region that received considerable attention (and was one of the reasons for choosing this area for GATE) was tropical cyclone “seedlings”, or cloud clusters. Several papers have been published that have analyzed cloud clusters that developed or moved through the GATE observational array. Martin and Schreiner (1981) noted that cloud clusters during GATE were separated by 5°-7° intervals and tended to split in two once they moved from the African continent over the ocean. The northerly clusters were African easterly waves, while the southern clusters were associated with a low-level monsoon circulation. Figure 50 shows the areas were cloud clusters were most common during GATE. Jenkins (1995) used GATE data to examine the temperature structure of African easterly waves.

Ott et al. (1991) used soundings from GATE ships to understand outbreaks of Saharan dust from the continent over the Atlantic basin. As discussed in Chapter 5, these dust outbreaks are now known to play a critical role in tropical cyclone intensification (weakening) and genesis (storms do not form when embedded in these layers). Figure 51 shows a cross section of one of the dust outbreaks examined in Ott et al. (1991). Dust outbreaks are now routinely tracked via satellite imagery – they can sometimes move all the way across the Atlantic and even into the eastern Pacific.
Figure 50. Cluster 3 hour occurrences over the primary area for the total period of GATE. All clusters are included. The resolution is $2^\circ \times 2^\circ$ (From Martin and Schreiner 1981, Copyright American Meteorological Society).

Figure 51. Zonal vertical cross section of Saharan Air Layer (SAL) thermodynamic structure along $15^\circ N$ at 1200 UTC 5 September 1974. Potential temperature ($2^\circ C$ contour interval) is plotted. The SAL is shaded (From Ott et al. 1991, Copyright American Meteorological Society).
Tropical Oceans Global Atmosphere (TOGA) – Coupled Ocean Atmosphere Response Experiment (COARE)

Another acronym that deserves a spot in the acronym hall of fame. The TOGA-COARE took place in the tropical West Pacific, known as the “warm pool”. The warm pool contains the warmest ocean SSTs in the world. It also experiences the strongest precipitation and atmospheric latent heat release globally. It is not hard to imagine that this area is critical both to understanding the general circulation of the Earth and the relationship of the warm pool to the El Niño – Southern Oscillation (ENSO), for example.

The goal of the TOGA program is to predict the variability of the coupled ocean-atmosphere system on time scales of months to years. The COARE was designed to work toward this goal. The goals of COARE (Webster and Lukas 1992) are to describe and understand:

1) the principal processes responsible for the coupling of the ocean and the atmosphere in the western Pacific warm-pool system;
2) the principal atmospheric processes that organize convection in the warm-pool region;
3) the oceanic response to combined buoyancy and wind-stress forcing in the western Pacific warm-pool region; and
4) the multiple-scale interactions that extend the ocean and atmospheric influence of the western Pacific warm-pool system to other regions and vice versa.

It involved a 4-month period of high density observations in the warm pool region. This period was called the “Intensive Observing Period”, or IOP. It took place during the months of November 1992 through February 1993 and involved a wide range of observing platforms and systems, including ships, buoy arrays, satellites, high density soundings, and aircraft measurements. The high resolution/quality datasets were then used to improve air-sea interaction and boundary-layer parameterizations in numerical models of the ocean and atmosphere, as well as to validate coupled models. The domain of the TOGA COARE is shown in Figure 52.

Important Findings of TOGA COARE

As the experiment was completed quite recently (1992), results using the high-quality data are still being analyzed and published. Several papers were published in the 1990s that described new datasets or the environment in which the IOP occurred. A paper published in 1999 (Curry et al. 1999) described the creation of a high-resolution dataset of surface fluxes of heat, precipitation, and momentum for the IOP. Other papers have documented the climatology (Hennon and Vincent 1996), lightning occurrence (Orville et al. 1997), and ship (Petersen et al. 1999) and aircraft (Yuter et al. 1995) operations during the IOP.

Important scientific findings from COARE will arise as the new datasets are analyzed.
Figure 52. The TOGA COARE Intensive Flux Array (IFA), where the highest density observations were taken during the Intensive Observing Period (IOP).

**Tropical Cloud Systems and Processes (TCSP) Mission**

The TCSP mission is sponsored by NASA. The focus of the study is to learn about the dynamics and thermodynamics of precipitating cloud systems and tropical cyclones through intense observation by NASA aircraft and satellites. The first phase of TCSP was based in San Jose, Costa Rica from July 1 – 27, 2005. In coordination with NOAA Hurricane Research Division (HRD) P-3 research aircraft, NASA flew 12 missions with their stratospheric jet (ER-2) into Hurricanes Dennis, Emily, Tropical Storm Gert, and an eastern Pacific mesoscale complex that may have evolved into Tropical Storm Eugene. ER-2 missions were flown into storms with P-3 aircraft below, all taking simultaneous measurements of a wide variety of atmospheric variables.

Post-storm TCSP research will address the following areas:

1) Tropical cyclone structure, genesis, intensity change, moisture fields, and rainfall;
2) satellite and aircraft remote sensor data assimilation and validation studies pertaining to development of tropical cyclones; and
3) the role of upper tropospheric /lower stratospheric processes governing tropical cyclone outflow
Review Terms

GATE

Cloud clusters

Warm pool

TOGA COARE

IOP

TCSP

References


Hennon, C.C., and D.G. Vincent, 1996: Climatology during the TOGA COARE Intensive Observing Period (November 1992 – February 1993) compared to a recent climatology. Purdue University, 95 pp.


Chapter 12
Cumulus Parameterizations

One unique aspect of the tropics is that a large percentage of precipitation there is convective in nature. That is, the precipitation is the result of deep cumulonimbus clouds such as those found in the ITCZ and mesoscale convective complexes. Stratiform precipitation does exist in the tropics, but to a much smaller extent than the extra tropics.

Numerical models have been used since the middle of the 20th century to predict the evolution of the atmosphere. Computer models have an advantage in the middle latitudes in that the important features of the atmosphere (long-waves, mid-latitude cyclones) are of large enough scale to be resolved in the models. In the tropics, the temperature gradients are very small and organized systems take the form of barotropic-like waves that are driven by latent heating within convective cores. Recall that mid-latitude systems extract their energy from the high temperature gradients found there.

The problem with tropical systems is that most numerical models cannot resolve (at least, horizontally) the small-scale convection that drives the larger-scale flows. This is true both in space and time. Convection typically lasts from a few minutes to at most a few hours. Even higher resolution mesoscale models (with horizontal resolutions on the order of 10 km and time steps of an hour or two) cannot explicitly model convection.

This presents a problem. We know that there are important feedbacks between convective activity and the larger-scale flow. For example, tropical cyclogenesis requires intense latent heating (created by towering cumulonimbi) near the core. This heating then lowers the surface pressure and creates more large-scale convergence into the system. If our computer model cannot “see” the latent heating going on in the core, then it will not be able to correctly model the increase in synoptic-scale convergence.

So does that mean we are stuck? Well, for now there are not powerful enough computers (nor the physical knowledge!) to explicitly model convection in an operational computer model. But we can create parameterizations for convection. The principal assumption in using convective parameterizations is that statistical properties of chaotic subgrid-size processes can be deduced from a knowledge of the resolved variables. In other words, if we can accurately describe the net effect that an area of small-scale convection is having on the grid-point scale of the model, then we don’t really need to know the gory details of what is happening at the smaller scales.

For example, let’s assume that there are three groups of students taking a tropical meteorology exam. Can you predict which group is going to do the best? Of course not! You need to know a little bit about each group. Let’s say the first group is a bunch of philosophy majors that have never had a science class in college. The second group is students who are currently taking an introductory atmospheric science class. Finally, the third group is doctorate students studying tropical meteorology and dynamics. Now can you predict which group will do best? Probably. This is the first step in making a cumulus parameterization – determining what type (deep, shallow, medium) of convection is occurring within a grid box and how much of the grid box is covered with each type. This is done by statistically analyzing the large-scale forcings. For
example, it has been found that deep (up to near the tropopause) convection most frequently occurs where there is a lot of low-level moisture convergence in a particular grid box.

Now that we know (to a rough approximation) the type(s) of convection (if any) occurring, we need to predict how the convection will impact the grid-point scale variables such as temperature, cloud cover, humidity, etc. This is analogous to predicting what the class average will be for the three groups of students. Without much study you can probably conclude that group 3 > group 2 > group 1 – but by how much? Well, you can be certain that the class averages for all three groups will vary by trial. One year group 1 may average 36/100, the next maybe 42/100. Let’s say that you do this experiment 5 times a year for 30 years. Then you will have 150 “trials”. Based on your trials you can determine a set of statistics for each class. For example, out of 100 points, group 3 averaged 96.5 with a standard deviation of 4.5. Assuming the test scores have a normal distribution (bell shaped), a mean and standard deviation completely describe the dataset.

So with a lot of analysis, you can relate the “net effect” that a certain type and extent of convection will have on nearby model grid points. We don’t care about the details. If one of the students in group 1 (the philosophy students) is a scientific genius who didn’t major in meteorology because his parents wanted him to be the next Kant (some philosophy dude I remember from college – look it up) and he scores a 99/100, then we have an anomaly. Sure, he’ll have some impact on the class mean, but overall his impact will be smoothed out. Convective parameterizations are not designed to account for the anomalies in nature, just as statistical forecast schemes will never be able to forecast an extreme event.

The following two sections describe two of the major cumulus parameterizations: the Kuo (1974) scheme and the Arakawa-Schubert (1974) scheme. Since there introduction there have been many improvements and modifications to each but we’ll leave that for the numerical weather prediction class. In addition, one can find a wide range of different convective parameterizations floating around out there, many being used in the global and mesoscale models run everyday. The descriptions that follow are rather simplified – more details (extreme details in some cases) can be found in Emanuel (1994).

**Kuo Convective Parameterization**

This scheme is based off of the Charney and Eliassen idea of convective instability of the second kind (CISK). The convection in Kuo’s parameterization requires that the large-scale circulation replenish the boundary layer with moisture. This maintains the conditional instability needed to keep the convection going.

The scheme is activated within a computer model whenever the atmosphere (as determined by the model) is unstable to deep convection. One of the assumptions of the model is that the precipitation rate is directly balanced by the net change of water in the column with time:

\[ \int_0^z \frac{\partial \rho q}{\partial t} \, dz = \alpha P \]
where $\bar{\rho}$ = average density of water in column, $q$ = specific humidity, $P$ = precipitation, and $\alpha$ is a dimensionless parameter needed to allow the scheme to produce realistic results.

There are several problems with the Kuo scheme:

1) Kuo predicts that the water vapor mixing ratio remains constant beneath the cloud as precipitation occurs – this is not matched by physical reasoning. The boundary layer should moisten as precipitation fall through.
2) Observations show that the atmosphere dries out during precipitation – Kuo predicts no such change
3) It cannot produce a realistic moistening of the atmosphere
4) The scheme does not depend on available convective energy to be activated.

The last point is rather troubling. When the scheme does not activate when the amount of energy available would spawn convection in nature, the energy will build up for a number of time steps before finally being released at the model grid points. You are left with a massive, unrealistic blow up of convection. These so-called “grid-point storms” are symptomatic of a deficient convective parameterization and negatively impact the larger-scale forecast.

Nevertheless Kuo’s scheme became very popular, primarily because it did a sufficient job overall and is computationally cheap.

**Arakawa and Schubert Convective Parameterization**

In the very same journal in which Kuo presented his parameterization, Arakawa and Schubert (AS) presented an alternative convection parameterization. Their scheme is called a “mass flux parameterization” because the heights of the cumulus clouds depend on the amount of mass convergence in the low-levels. The AS parameterization is more computationally intensive than the Kuo scheme, but has a stronger physical basis.

For each model grid box, there can be different cloud heights. Each type of cloud height has a different “entrainment rate” (the rate at which unsaturated outside air is taken in by the cloud) – shallow cumulus clouds have a large entrainment rate, while deep cumulus have a low entrainment rate. If the mass flux exceeds the entrainment, the convection will grow – otherwise the convection will cease.

Although more physically sound than Kuo, the AS scheme does come with some concerns. Convection cannot initiate in any atmospheric level except the planetary boundary layer. This restriction has been relaxed in subsequent versions of the model. Also, the cloud type must be determined a priori (first, before the parameterization is run) – this done automatically by analyzing the large-scale environment for things like moisture convergence. Finally, the entrainment rates the AS scheme uses have been shown to be very sensitive. So if they are off by just a bit, it could have a significant impact on the model forecasts.
For more information on convective parameterizations like the Kuo and AS schemes, see Emanuel (1994) or http://www.ecmwf.int/newsevents/training/rcourse_notes/pdf_files/Conv_mass_flux.pdf

**Review Terms**

Parameterizations

Kuo scheme

Arakawa and Schubert scheme

**References**


Section III: Tropical Motion and Oscillations

Chapter 13
General Circulation

The general circulation consists of the global flow averaged over a long enough time to eliminate variations due to weather systems but a short enough time to account for seasonal and monthly variations. It deals with the distribution and evolution of such things as temperature, humidity, precipitation, wind, and other variables.

George Hadley was an 18th century British climatologist who first proposed that there existed a thermally driven circulation that was responsible for the permanent trade winds. He theorized that differential heating between the tropics and poles forced poleward flow at upper levels and a return flow at the surface toward the Equator. The Coriolis force was responsible for turning those winds, resulting in easterly trades in the tropics. This circulation is still called the Hadley Circulation today, although it is not a hemispheric circulation that Hadley envisioned but rather one limited to the tropics.

Forcing of the General Circulation

There is a radiative imbalance between the tropics and the poles. That is, incoming solar energy is strongly dependant on latitude, but the outgoing longwave is only weakly dependant on latitude. Figure 53 below shows that there is a net surplus of energy in the tropics and a net deficit at the poles.

![Radiation Budget](image)

*Figure 53. Annually averaged radiation budget by latitude.*

Therefore, the tropics experience a net warming and the poles a net cooling. This sets up a horizontal temperature gradient between the two regions. Baroclinic instabilities (e.g. mid-latitude cyclones) develop and transport some of that heat poleward. In addition, baroclinic...
disturbances also convert potential energy (from the temperature imbalance) into kinetic energy, maintaining atmospheric flow that is being dissipated by friction.

The general circulation can also be thought of as a “heat engine”, where energy is absorbed at warm temperatures in the tropics (mostly due to latent heat from water evaporation) and given off at cool temperatures in the extratropics. So the thermal imbalance converts potential energy into kinetic energy through the convective overturning.

Figure 54 shows the cross-section of the general circulation model. The following sections will discuss the trade winds and equatorial low (ITCZ). A unique convergence zone called the south Pacific convergence zone (SPCZ) will then be presented. This will be followed by a brief discussion on tropical monsoon circulations, which are seasonal phenomena connected to the general circulation but affect a smaller region.

![Figure 54. Schematic of the 3-cell general circulation model (from http://web2.uwindsor.ca/courses/biology/macisaac/55-437/lecture1/HADLEY.JPG).](http://web2.uwindsor.ca/courses/biology/macisaac/55-437/lecture1/HADLEY.JPG)

**Trade Winds**

As shown in Figure 54, the trades are easterly winds that flow from the latitudes of the sinking branch of the Hadley cell (~ 20°-25°) toward the Equator. The winds are result of a horizontal pressure gradient force set up from the subtropical high pressure belt and the Equatorial low (ITCZ). The winds do not flow in a straight manner to the Equator because they are turned to the right (NH) and left (SH) by the Coriolis force.
The speed of the trades vary with location, but generally average 10-15 kt. They are generally steadier over the oceans, where surface friction is less a factor. The trades are affected by intraseasonal oscillations such as El Niño. During such an event, the easterly trades in the central Pacific drastically weaken or even reverse direction. This will be discussed more later in the course.

The trade winds gradually weaken as they approach the Equatorial low, where convergence with the opposing hemisphere’s trades occurs. Surface winds are very weak – this area is called the “Equatorial low doldrums”, and represents the location of the Intertropical Convergence Zone (ITCZ).

**Intertropical Convergence Zone (ITCZ)**

The ITCZ is defined as a narrow (~5°) band of east-west vigorous deep convection that meanders back and forth across the equator. The convection is not continuous, but one can usually clearly see the ITCZ on a satellite image. Figure 55 is one such image.

![ITCZ from GOES-West](image)

*Figure 55. The ITCZ from GOES-West.*

Following is a list of a features and characteristics of the ITCZ:

- generally located near maximum SSTs
- heavy precipitation; usually precipitation exceeds evaporation by a factor of 2 or more, even though this area of the world has the highest evaporation rates found anywhere
- undergoes large temporal and spatial variations; meanders N-S more over land than water
- in some locations (e.g. Africa, Western Pacific), easterly wave disturbances are tied to the ITCZ
• other variables, such as outgoing longwave radiation (OLR), are well-correlated with the ITCZ since it contains deep, active convection

The ITCZ is the primary producer of precipitation in the deep tropics. Its seasonal migration brings marked wet and dry seasons for many tropical areas, the most famous being the African Sahel region and the Asian monsoon. This shift is seen most clearly in a time-latitude diagram, called Hovmöller diagrams. Figure 56 shows 3 such diagrams for three separate longitudes: 100°E (south Asia to the north, ocean to the south of the equator), 130°E (east Asian coast to the north), and 145°E (mostly ocean). The variable plotted is OLR – the shaded areas represent OLR minimum, which correspond to cold cloud tops and thus deep convection. The x-axis is month, moving from left to right. Note that at all three longitudes, the ITCZ moves northward with the NH summer season. The shift is most pronounced at 100°E, where the Asian landmass promotes a much more northerly progression of the ITCZ. Also, observe that the ITCZ moves into the southern hemisphere during its summer, although it stays closer to the equator than the NH.

Figure 57 is another set of Hovmöller diagrams but for locations far out into the Pacific Ocean (160°E, 175°E, and 155°W). Three things to note from these plots:

1) there appears to be a “double ITCZ”, especially near 160°E;
2) the ITCZ does not meander N-S nearly as much over the ocean; and
3) there is another linear band of convection that shows up at around 20°S and 155°W. This is the South Pacific Convergence Zone (SPCZ), which will be discussed next.

Theories on ITCZ Cause and Maintenance

The currently accepted theory regarding the ITCZ is that it is a thermally-direct (meaning warm air is rising, cool air is sinking) planetary-scale circulation due to the low-level convergent winds (trade winds) forced by SST gradients. In other words, the intense heating of the sun warms the ocean surface most near the equator, which destabilizes the atmosphere, causing deep and persistent convection. This is the driver of the general circulation. There have been other theories set forth, however:

1) Charney and collaborators think that the ITCZ is a synoptic-scale phenomena that is caused by a CISK-type forcing. Recall that CISK theory is a cooperation between the localized convection and the large-scale (e.g. trade winds)

2) Others postulate that the ITCZ is located where the maxima of vertically-integrated moist static energy (sensible heat + latent heat + potential energy) is found.

There are several problems with each theory. The CISK theory neglects the sensible and latent heat fluxes, relying on vast storages of potential energy that observationally are shown not to exist. Also, potential energy exists in many other environments that do not create an ITCZ like feature (e.g. mid-latitude continental areas).
Figure 56. Time-latitude plots of outgoing longwave radiation (OLR) for a) 100°E, b) 130°E, and c) 145°E.
Figure 57. As in Figure 56, except for d) 160°E, e) 175°E, and f) 155°W.
Although the energy theory (#2) correctly positions the ITCZ where observations of moist static energy maximize on long-term time scales, it fails to do so with shorter (< 1 year) time scales.

**South Pacific Convergence Zone (SPCZ)**

In early geostationary satellite analyses of the Western and Central Pacific, a secondary convergence zone was identified. It originated in the tropical warm pool and was oriented in a northwest to southeast diagonal fashion, terminating in the south Pacific. This area of semi-permanent convection was called the SPCZ.

Figure 58 shows a schematic diagram of the SPCZ, and its location relative to the ITCZ and the Pacific basin. It is formed by the convergence of cool mid-latitude southwesterly winds with warm, moist trade winds from the eastern Pacific. As with the ITCZ, the SPCZ meanders but is not connected with the seasonal cycle as strongly. For example, it may move with a period of weeks to interannually. During El Niño events, the SPCZ will be displaced north and east of it’s average position.

Unlike the ITCZ (which is forced and maintained entirely in the tropics), the SPCZ is thought to be maintained through tropical-higher latitude interactions. This includes such phenomena as tropical and sub-tropical cyclones moving poleward and southern hemispheric mid-latitude storms moving toward the tropics from higher latitudes. The location and strength of the SPCZ can also be affected by SST gradients and the movement of the subtropical high pressure center in the Eastern South Pacific. The SPCZ is climatologically strongest during the month of January.
Monsoon Circulations

A monsoon is defined as a seasonal reversal in the wind circulation. It is commonly thought of as a period of extreme rainfall over some tropical area, but that is really only half of the story. It is actually a planetary-scale phenomenon that affects areas between about 10°S to 35°N.

Monsoon circulations are rather ubiquitous, occurring over most continents. However, the most famous and vivid example of a monsoon circulation is the Indian summer monsoon, or the Asian monsoon.

Monsoon Lifecycle

1) The first signs of the monsoon onset occur in late April off the east coast of Madagascar (~20°S, 60°E), where a band of strong southeasterly winds blows at low levels (3000 – 7000 ft). These are thought to be caused by cold air outbreaks over the southern Indian Ocean.
2) Over the next few weeks, this flow progresses northward toward the equator; by late May, this “low-level jet” reaches the African coast near the equator at Kenya/Somalia.
3) The high plateau of east Africa turns the winds clockwise toward the Arabian Sea.
4) Onshore flow sets up from about late-May through September over India/Arabian Sea. This flow brings very warm/moisture laden air into the Asian continent. Rainfall is enhanced by topographic lift due to higher elevations to the north.

5) Flow reverses in fall as the interior Asian continent becomes cold. This reverses the pressure gradient and causes offshore flow, bringing cold, dry air into India from the north. Precipitation shuts down until next summer.

Precipitation amounts from the wet phase of the monsoon can be staggering. Some locations in India receive 420” of rain (yes, that’s 35 FEET) in a single summer.

Review Terms

General circulation

Trade winds

ITCZ

SPCZ

Monsoon circulation
Chapter 14
Atmospheric Waves in the Tropics

Equatorial Waves

Much of the convection seen along the ITCZ originates from equatorial waves, which are disturbances that propagate along it. They derive their energy from the conversion of diabatic heating into kinetic energy.

Equatorial waves have been observed to move 8-10 m/s toward the west. Their period is 4-5 days.

It is believed that equatorial waves play a role in enhancing the large-scale convergence into the ITCZ. A large area of deep cumulonimbi convection creates a regional heating source, which induces larger-scale upward motion in a CISK-like process. This will lower the surface pressure, inducing a stronger low-level convergence into the area. One requirement of CISK is that the tropical atmosphere contains a lot of CAPE. Observations have shown that this requirement is usually not met. Therefore, alternative hypothesis have been put forward, including Wave-CISK and Wind Induced Surface Heat Exchange (WISHE).

Equatorial waves have a deep layer of low-level convergence, as shown in the Figure 59:

![Figure 59. Divergence profile of a typical equatorial wave.](image)
This implies that very dry mid-level air is entrained into the cumulonimbi clouds. Usually, this will entail a killing off of the convection as strong evaporative cooling creates strong downdrafts. However, the core updrafts of the equatorial waves are typically sheltered from the entrainment, allowing the waves to persist for many days.

**African Easterly Jet**

During the Northern Hemisphere summer, strong heating of the Saharan region in North Africa creates a situation in which the usual north-south horizontal temperature gradient is reversed (i.e. temperature increasing toward the north). The thermal wind relation calls for a strong change in geostrophic wind with height over such an area of steep horizontal temperature gradient (Figure 60). Observations show an easterly jet stream in this region during the summer. This regional jet stream is called the **African Easterly Jet (AEJ)**.

![Figure 60. Schematic of the thermodynamic situation that leads to the African easterly jet.](image)

A cross section of the zonal wind in the AEJ (Figure 61) shows a jet maximum located around 650-700 mb, centered around 16°N.

**African Wave Disturbances (Easterly Waves)**

One important impact of the AEJ is the creation of African wave disturbances (AWDs). These are synoptic-scale features that form within the cyclonic shear zone on the south side of the AEJ. They have a phase speed of about 8 m/s and a period of 3.5 days. Over the African continent, these disturbances tap the AEJ for most of their energy. After they move offshore (where the AEJ disappears), African waves must transition to a diabatic heating energy source.
This transition happens successfully most often in the late summer/autumn, when sea surface temperatures (SSTs) off the west coast of Africa are warm enough to sustain the intense convection of AWDs. As the disturbance continue their westward march, they may form into tropical cyclones, assuming other favorable atmospheric conditions are in place (e.g. low vertical wind shear). As shown in the satellite image below (Figure 62), AWDs may continue on into the Pacific basin – sometimes the low-level convergence will do so without the existence of larger-scale convection.

AWDs are also commonly called “African Easterly Waves”. It is believed that a large percentage (> 50%) of Atlantic TCs can trace their heritage back to an easterly wave. In recent times, some believe that an even higher percentage of Atlantic (and Pacific storms) come from easterly waves.
Figure 62. Satellite image showing a train of African easterly waves. The wave in the central Atlantic appears to be developing into a mature tropical cyclone.

**Review Terms**

Equatorial waves

African easterly jet

African wave disturbances

Easterly waves
Chapter 15
Madden Julian Oscillation (MJO)

Introduction

The MJO is an intraseasonal oscillation- defined as one that is longer than synoptic-scale (2-5 days), but shorter than a season (~90 days). Usually, they are classified as having periods of 7-70 days.

Background

Although several higher frequency oscillations have been identified (e.g. Schrage and Vincent 1996), the primary tropical intraseasonal oscillation is the MJO.

Work in the early 1970s at the National Center for Atmospheric Research (NCAR) by Roland Madden and Paul Julian led to the discovery of a 40-50 day oscillation in the zonal wind in the tropics. This oscillation came to be known as the Madden-Julian Oscillation, or MJO (Madden and Julian 1971, 1972).

Essentially, the MJO is a wave in the atmosphere, centered in the tropics, that moves easterly around the globe. Some characteristics of the MJO:

• Usually wavenumber 1. This means that the oscillation is usually symmetrical and circular in shape.
• The amplitude of the wave varies. Sometimes the MJO can be detected as far as 20°-30° away from the Equator. The strongest amplitude (and slowest propagation speed) usually occurs between 60°E and the Dateline (180°)
• The MJO can be identified by a maximum in the upper level (~200 mb) divergence field. It is frequently accompanied by convection, except east of the Dateline where the convection becomes uncoupled from the circulation.
• Average phase speed ~10 m/s (~5 m/s between 60°E and the Dateline)
• A dipole in the upper level divergence exists between the Indian Ocean (60°-120°E) and the Central Pacific/SPCZ region. In other words, when one area shows a divergence maximum, the other shows a convergence maximum.
• The upper level divergence field moves around the globe. However, the convection tends to be maximum in the Indian Ocean, dissipate, and then reform near 160°E.

A schematic diagram illustrating this last point is shown in Figure 63. This figure shows the sequence of shifts in the Outgoing Longwave Radiation (OLR) anomalies – where one would observe convection due to the MJO. Anomalies in areas 1 and 3 tend to be out of phase. Thus, if convection was present in area 1, clear skies would persist in area 3.
Observations and Structure of the MJO

Figure 65 is a schematic diagram of the MJO in the vertical. The arrows represent the departure from the mean Walker circulation, and time progresses downward from the top of the figure. Note that as the MJO perturbation approached from the west, an observer would first experience an anomalous easterly wind (e.g. strengthening of the trade winds). After the convection passes, winds have a westerly anomaly component. This will significantly weaken or even reverse the trade winds. Note also the features of the convection as it disappears in the central Pacific (but the convergent/divergent pattern continues on its eastward trek).

Perhaps the clearest manifestation of MJO appears in a time lapse figure of OLR. Figure 64 shows a time-longitude diagram (called Hovmoeller plots) of OLR for a 6-month period averaged from 10°S to 2.5°N. Note how convective areas (negative or dark OLR anomalies) move from west to east across the domain. The period between negative OLR anomalies is about 40 days (same period as the MJO).

Importance of MJO

So what is the big deal about the MJO? The MJO is the primary regulator of intra-annual tropical weather. Although it is most obvious in the Indian and western Pacific Oceans, the MJO affects weather throughout the tropics. It drives variations in many meteorological and oceanic variables, including wind, SST, cloudiness, and rainfall. Recent work has found a connection between the MJO cycle and tropical cyclogenesis, especially in the eastern Pacific. For example, Maloney and Hartmann (2000) found that
Figure 64. Time-longitude section of the OLR anomalies for the MJO-filtered band averaged for the latitudes from 10°S to 2.5°N. The zero contour has been omitted. Light shading for positive anomalies and dark shading for negative anomalies (From Wheeler and Kiladis 1999 – Copyright American Meteorological Society).

during periods of westerly equatorial 850 mb wind anomalies over that region twice as many hurricanes and tropical storms form. Also during that time, systems that do form are stronger than in easterly anomaly periods. Hurricanes in westerly wind anomaly periods outnumber those in the easterly period by a 4 to 1 margin. Among the factors cited by Maloney and Hartmann that explain this is enhanced cyclonic vorticity, lower vertical wind-shear, and enhanced convection spawned by the MJO wave. Figure 66 shows the differences in frequency and intensity of E. Pacific systems as a function of the MJO phase. Phase 2 (6) represents easterly (westerly) 850 mb wind anomalies.

Simulation of MJO in Global Climate Models

It has been a big challenge to accurately represent the MJO in model experiments. The Community Climate Model Version 3 (CCM3) simulated a weak MJO that tended to move in the opposite direction (westerly) of observations in the Indian Ocean. However, recent work by Zhang and Mu (2005) have resulted in a simulated oscillation that closely matches the 850-mb zonal wind, precipitation, and OLR observations of a real MJO. However, the time period of the simulated MJO is shorter (~30 days vs. 40-45 in reality) and the spatial extent of precipitation is smaller than observed. The main improvements were made in the convection scheme of the CCM3.
Figure 65. Schematic depiction of the time and space (zonal plane) variations of the disturbance associated with the 40-50 day oscillation. Dates are indicated symbolically by the letters at the left of each chart. Regions of enhanced large-scale convection are indicated schematically by the cumulus and cumulonimbus clouds. The relative tropopause height is indicated at the top of each chart. (From Fig. 16 of Madden and Julian (1972b) – Copyright American Meteorological Society).
Figure 66. At left, number of hurricanes and tropical storms as a function of MJO phase for the eastern Pacific Ocean hurricane region during May-Nov 1979-95. At right, average strength (kt) of hurricanes and tropical storms as a function of MJO phase for the same period. Error bars represent 95% confidence limits (From Maloney and Hartmann 2000 – Copyright American Meteorological Society).

References


Chapter 16
El Niño – Southern Oscillation (ENSO)

History

1877 – Drought and famine in south Asia – monsoons fail

1891 – Señor Dr. Luis Carranza (President of Lima Geographical Society) writes small article about a counter-current flowing from north to south off coast of Peru

Paita (port of Peru) sailors call current “El Niño” because it appeared immediately after Christmas

Carranza noted that great rains fell over the barren desert adjacent to the cold ocean during this time. These years were known as “años de abundancia” (years of abundance) – pasture replaced deserts. However, marine life and birds left, at least temporarily.

1897 – Hildebrandsson notices that atmospheric pressure oscillations in Sydney Australia are out of phase with those in Buenos Aires Argentina.

1899 – Monsoons fail in India again

1902 – Lockyer and Lockyer (father and son) confirmed Hildebrandsson’s observation and established the period of oscillation to be about 3.8 years and nearly global in extent.

1904 – Sir Gilbert Walker becomes Director-General of Observatories in India. He wants to be able to predict interannual monsoon variability.

- Established the existence of a “Southern Oscillation”, a see-saw in pressure between the Indian Ocean and eastern tropical Pacific. Noticed that they were out of phase and sought to connect this observation to monsoon predictability.

1923 – 1937 – Walker and colleagues publish a series of papers in which the Southern Oscillation is shown to be correlated with changes in rainfall patterns and wind fields over the tropical Pacific and Indian Oceans – failed to find a reliable predictor of monsoon activity

1940’s – 1950’s – Interest in the phenomenon wanes as scientists are unable to physically explain Walker’s observations

1957 – 1958 – Better Sea Surface Temperature (SST) data shows a large correlation between SSTs and Southern Oscillation. Jacob Bjerknes (UCLA) proposes that the coincidence between oceanographic and meteorological data is not unique to 1957-1958 but occurs interannually.

1969 – Bjerknes proposes physical explanation – sinking air over the cool eastern Pacific waters flows westward and rises in vast cumulonimbi clouds in the west. Names this west-east
circulation “**Walker Circulation**”. SST gradients force this circulation. If SST gradients weaken, convective cells move eastward into the central and eastern Pacific.

Bjerknes thought that interactions between the ocean and the atmosphere drove the Southern Oscillation. An initial change in the atmosphere (e.g. a slight weakening of the trade winds) will lead to a slight warming in the eastern Pacific, which then would lead to an even greater weakening of the trades, etc…. He proposed “a never-ending succession of alternating trends by air-sea interaction in the equatorial belt” as the cause of the Southern Oscillation. Bjerknes was unsure of the physical mechanisms that caused a change of phase (El Niño to La Niña or vice versa).

**Background**

The **southern oscillation** refers to the pressure fluctuations in the tropics with centers of action in the western Pacific/eastern Indian Oceans and the southeastern Pacific. This is shown in the Figure 67 below. Note the strong negative (positive) correlation in sea level pressure centered over 120E (120W).

![Image of Figure 67](image)

*Figure 67. Correlations of annual mean (May through April) of sea level pressure with the SOI [southern oscillation index] for 1958-1998. Values greater than 0.6 are hatched and those less than -0.6 are stippled. (From Trenberth and Caron 2000).*

**El Niño** is the phase of the southern oscillation in which there is a high (low) pressure anomaly in the western (eastern) Pacific that coincides with anomalously warm SSTs and heavy rainfall in the east, a relaxation of the easterly trades, and an anomalously deep thermocline in the eastern Pacific. A typical El Niño lasts anywhere from 12-18 months, but there is much variation. The 1990-1995 El Niño is one example. There is a 2-7 year period on average between El Niño events.
**La Niña** is the opposite phase of the southern oscillation, characterized by low (high) pressure in the western (eastern) Pacific, anomalously cool SSTs across much of the eastern and central Pacific, and a strengthening of the easterly trades.

**Determining the phase of the Southern Oscillation**

There is no one definition of when there is an El Niño, La Niña, or something in between. Each event is distinct from each other. There are a couple of methods that are used to determine the existence and strength of these events. One focuses on the Southern Oscillation itself, the other measures differences of SST in the ocean.

The **Southern Oscillation Index**, as its name implies, tracks the see-saw in pressure between the eastern Pacific/Indian Ocean and the central Pacific. It is the difference between deseasonalized (seasonal trends removed), normalized sea level pressure (SLP) anomalies over Tahiti and Darwin, Australia. Although one can use monthly means to determine trends in the SOI (shorter periods have contamination from higher frequency weather patterns), it is usually analyzed as a 3 or 6-month running mean. Figure 2 below shows the SOI as calculated by the Australian Bureau of Meteorology. You should be aware that there are several methods for calculating the SOI – it seems that the Australian method is scaled by a factor of 10 (see [http://www.bom.gov.au/climate/glossary/soi.shtml](http://www.bom.gov.au/climate/glossary/soi.shtml) for more information):

\[
SOI = 10 \frac{P_{\text{diff}} - P_{\text{diffav}}}{SD(P_{\text{diff}})}
\]  

where:

- \(P_{\text{diff}}\) = (Average Tahiti MSLP for month) – (Average Darwin MSLP for month)
- \(P_{\text{diffav}}\) = Long term average of \(P_{\text{diff}}\) for that month
- \(SD(P_{\text{diff}})\) = Long term standard deviation of \(P_{\text{diff}}\) for the month.

Using this method, the SOI from 1990 – 2006 is shown in Figure 68. Positive (negative) values indicate La Niña (El Niño) events.

![Figure 68. 6-month running mean of the SOI as calculated by the Australian Bureau of Meteorology. Dark (negative) shades indicate El-Niño events while lighter (positive) shades indicate La Niña events.](image-url)
There are other indices based on SST anomalies in the Pacific rather than pressure anomalies of the atmosphere. Figure 69 shows the averaging regions that are used to determine the phase of the ENSO cycle.

Two common indices from SST data are the Niño 3.4 (shown in Fig. 69 above) and the Trans-Niño Index (TNI), which is determined from normalized anomalies between the Niño 1+2 and Niño 4 regions. See Trenberth and Stepaniak (2001) or http://www.cgd.ucar.edu/cas/catalog/climind/TNI_N34/index.html#Sec5 for more details.

The **Multivariate ENSO Index** (MEI) is a sort of hybrid indicator of ENSO that accounts for both atmospheric and oceanic variables. The six variables are:

1. Sea-level pressure (P)
2. Zonal component of the surface wind (U)
3. Meridional component of the surface wind (V)
4. Sea surface temperature (S)
5. Surface air temperature (A)
6. Total cloudiness fraction of the sky (C)

A recent MEI plot is shown in Figure 70 below.

![Figure 70. The MEI from 1950 – 2006. From http://www.cdc.noaa.gov/people/klaus.wolter/MEI/mei.html](http://www.cdc.noaa.gov/people/klaus.wolter/MEI/mei.html)
Physical Explanation of ENSO

As mentioned above, Bjerknes was the first to connect the southern oscillation to changes in the ocean. He realized that there was a circular relationship between changes in each regime, but could not establish which came first. Due to the inherent complexities in understanding a physical system as large and diverse as ENSO, there is still no definitive explanation as to how the events unfold. But there is an understanding about the feedbacks involved between the ocean and atmosphere, and how ENSO evolves once a phase is set in motion.

The key to understanding ENSO occurred in the 1960s with the realization of the importance of the equatorial trapped Kelvin and Rossby waves. Although these waves occur both in the ocean and atmosphere, they move much more slowly in the ocean. Kelvin waves propagate eastward at 2-3 m/s, while Rossby waves move westward more slowly (0.6 – 0.8 m/s). Both carry energy and momentum gained from surface wind stresses.

Kelvin Waves

Kelvin waves are equatorially trapped. This means they are confined to a narrow, north-south region centered on the equator. The equator is called a wave guide. If you imagine a roller coaster attached to a rail but able to oscillate back and forth on either side, you get the idea what a wave guide is (the rail). The restoring force is changing sign of the Coriolis force on each side of the equator. A wave with a northward component will get turned back to the south (to the right) by the NH Coriolis force.

Kelvin and Rossby waves are forced by wind anomalies in the tropical Pacific. These usually take the form of westerly wind bursts, which were intense westerly wind anomalies that can signal the passage of a MJO or the onset of an El Niño event. Kelvin waves take a few months to propagate across the Pacific Basin. Once they reach the western coasts of Central and South America, they can travel pole ward along the coastlines (Figure 71).

Bjerknes Hypothesis

The Bjerknes hypothesis can explain the onset of a warm or cold ENSO phase but cannot explain how the transition between the two can occur. Essentially, it is a positive feedback loop with no counter-balances. Figure 72 is a schematic diagram of an El-Niño event within this framework. A modest change in either the equatorial SST distribution
or the strength of the easterly trades feeds back into strengthening the response. For example, a weakening of the trades allows for warm water to slosh east and a deepening of the thermocline there. This further weakens the west-east pressure gradient, further weakening the trades.

But questions remain: What stops the warming in the eastern Pacific? Why do El-Niño events last 12-18 months? Why do El-Niño events rapidly transition to La Niña events?

Delayed-Oscillator Theory

The response of the ocean to surface wind forcing is an important consideration in this theory. The delayed oscillator theory works as follows (see Figure 73):

1. The westerly wind anomaly (e.g. weakened easterly trades) deepens the thermocline near the Equator, but raises it about 3°-8° away from the Equator. Recall that higher thermoclines indicate colder SSTs.
2. The high thermocline anomalies propagate westerly in the form of Rossby waves. They have little or no signal in the SST field since the climatological thermocline in the western Pacific is too deep to have any noticeable effect at the surface.

3. The maritime continent reflects the Rossby wave back toward the east as a Kelvin wave. This wave raises the thermocline as it moves along. When it reaches the eastern Pacific, where the climatological thermocline is closer to the surface, SSTs begin to cool, beginning the transition to a La Niña.

There is about a 6 month time lapse between steps 2 and 3, which fits with the average 12-18 month El Niño event. See http://physicsweb.org/articles/world/11/8/8 for more information.

**Stochastic Theory**

This theory stands upon the premise that the coupled ocean-atmosphere system is actually stable (not vulnerable to perturbations) and that ENSO events are triggered by random forcing from the atmosphere. This theory is attractive because it suggests that ENSO events should be irregular in both length and frequency, which matches what is observed.

![Figure 73](enso_73.png)

**Figure 73. Important features of the delayed-oscillator theory.** 1) Westerly wind anomaly creates westward moving Rossby waves, 2) Rossby waves advect anomalously shallow thermocline westward, 3) Rossby waves reflected off Maritime continent, return eastward, 4) Kelvin waves raise thermocline in the east, concluding warm phase and initiating La Niña event.

**ENSO Teleconnections**

**Tropical Cyclone Frequency**

As discussed in Chapter 2, there are a number of necessary (but not sufficient) environmental conditions that must be in place in order for tropical cyclogenesis to occur. Among those six (6) conditions was low values (< 10 m/s) of vertical wind shear (generally between 850 mb and 200 mb), especially over the center of the system. High vertical wind shear destroys the vertical thermodynamic structure of the developing system. In other words, the localized heating produced by the organizing convection is displaced away from the center of the developing
circulation. This severely limits additional inflow and thus strengthening of the primary circulation.

In the Atlantic basin, there is evidence of a **multi-decadal oscillation** in SST. This 15-40 year variation in SST is thought to be a driving factor in the number and severity of tropical cyclones in the basin (Goldenberg et. al, 2001). Since 1995 (and especially in 2005), there has been a noticeable increase in Atlantic basin activity, with the exception of the 1997. As seen in Figs. 2 and 4, that year was a strong El-Niño year. Could El-Niño be a limiting factor in the number of storms in the Atlantic?

The answer is yes. The ENSO impacts tropical cyclone activity in all basins. Some basins experience a change in frequency (e.g. Atlantic), while others show a shift in the genesis location of tropical cyclones. The impacts of El-Niño on tropical cyclone events in each basin can be summarized as follows:

1. Australian region (90°E - 165°E): Reduced frequency. The action shifts east of 165°E as the convection follows the warm SST anomalies. Also, the Australian monsoon trough, which provides a favorable environment for genesis, weakens during an El-Niño.

2. Northwest Pacific – Fewer tropical cyclones west of 160°E, more formations from 160°E to just east of 180°. Variations thought to be correlated with changes in the monsoon trough location and strength.

3. Northeast Pacific – More genesis near Hawaii (140°W to Dateline) and more TCs that track into the region the following year. There is no detectable correlation in the eastern portion (east of 140°W) of the region.

4. Atlantic – Reduced frequency. This region experiences an increase in the westerly winds at upper-levels (~200 mb). The increase in vertical wind shear noticeably decreases the number of hurricanes that the basin experiences.

5. Southwest Indian and North Indian – No detectable ENSO forced variations

Figure 74 summarizes the impacts that ENSO has on tropical cyclone activity around the globe.
United States Climate

ENSO cycles create changes in weather patterns outside the immediate tropical Pacific ("teleconnections"). The ENSO teleconnections are not forced by the SST anomalies themselves, but rather the influence that the SST anomalies has in shifting the tropical rainfall anomalies. This in turn alters larger-scale weather patterns outside of the region, leading to displacement of jet streams and storm tracks.

Precipitation

By far, the winter season shows the highest correlation with ENSO forcings. Figure 75 shows composite U.S. winter precipitation anomalies of El Niño events (top) and La Niña events (bottom) from the 20th century. The following observations can be made for El Niño precipitation teleconnections:

1. Southeast U.S. experiences much more precipitation
2. California coast experiences similar precipitation anomalies
3. The Pacific NW and Midwest are generally drier

The main reason is that the sub-tropical jet brings storms into the southeast and southwest U.S. more frequently during El Niño events. La Niña patterns are generally opposite of the El Niño pattern.
Temperature

U.S. wintertime temperature anomalies can be summarized as follows: warm in the north, cool in the south for El Niño, cool in the north and warm in the south for La Niña. Maximum temperature anomalies are approximately 2°-4° F.

ENSO Predictability

As with other meteorological forecasts, there are two approaches that are taken to forecast ENSO. **Dynamical models** are full fledged physics-based models that use proven physical relationships (e.g. equation of state, equations of motion, etc.) to predict the future state of a variable (in this case, SST). Although they theoretically can predict extreme events (since they are not bounded by statistics), dynamical models are computationally intensive. Furthermore, they cannot model small-scale features due to the resolution limitations, both spatially and temporally. **Statistical models** use past data to predict future behavior. They are simple, easy to use, and do not involve a lot of computations. Statistical models have difficulty in predicting extreme events since they are reliant upon a finite dataset of past events. Nevertheless, statistical models have shown great success in many aspects of meteorological forecasting, including tropical cyclone intensity and precipitation prediction.

Zebiak-Cane Dynamical Model

The first dynamical model to successfully forecast an ENSO event was created by Zebiak and Cane (1987). The ZC model is a coupled air-sea model, meaning that the ocean and atmosphere are able to “talk” to each other. The atmospheric component follow the Gill (1980) model – steady-state, linear shallow-water equations, equatorial beta-plane. The atmospheric circulation is forced by SST heating anomalies in combination with low-level convergence. The ocean component of the model is capable of reproducing thermocline depth anomalies, which is a critical inclusion when attempting to simulate the ENSO cycle. The upper 50 m of the ocean are able to respond to surface wind forcings. Ocean currents were initialized by spinning up the model with monthly mean climatological winds, and are further forced by surface wind stress anomalies. The time step of the ocean and atmospheric components are 10 days.

ZC published the results of a 90-year run of the model, initiated with an imposed westerly wind anomaly from 145°E - 170°W (after which there was no imposed external forcing). The area-averaged SST anomalies for the simulation are shown in Figure 76. Some important things to note:

1) Recurrence of warm (El Niño) events. Note that they are irregular (both in time space and in amplitude) and are pure manifestations of the coupled model.
2) Despite the irregularity, there is a favored period of 3-4 years in the oscillation.
3) Warm episodes tend to peak in June and last from 14-18 months.
4) Warm SST anomalies peak at 2°-3°C at the eastern boundary.
Figure 75. Wintertime U.S. precipitation anomalies during an average El Niño (top) and La Niña (bottom) event. From http://www.cdc.noaa.gov/ENSO/Compare/
Figure 76. Area-averaged SST anomalies for the 90-year model simulation. The solid line is NINO3 (5°N-5°S, 90°-150°W), and the dotted line is NINO4 (5°N-5°S, 150°-160°E). [From Zebiak and Cane (1987), Copyright American Meteorological Society].

Figure 77 shows SST observations for the period of 1921-1976 from Rasmusson and Carpenter (1982). Note that the SST variations are irregular, have a 3-4 year period, usually have maximum amplitude in summer, and peak between 2°C and 4°C for major ENSO events. This is a good indication that the ZC model is able to capture the important factors that drive ENSO events. They conclude that:

1) A necessary precondition for onset of El Niño is an above-normal equatorial heat content
2) All mechanisms essential to the ENSO cycle are contained with the tropical Pacific basin alone. This means that there are no external forcings, say from the mid-latitudes or elsewhere. However, as we saw above, this does not mean that ENSO itself does not drive teleconnections outside of the region.
3) Random forcing of unknown origin is not required. This is important – it means that it is indeed possible to predict El Niño events, even out to 1-2 years.
Since this landmark study, improvements to the model have been made. See http://rainbow.ldgo.columbia.edu/~dchen/forecast.html for more information.

Aspects of ENSO forecasts

Current ENSO forecasts involve the analysis of many dynamical and statistical models into a sort of “ENSO ensemble”. Figure 78 shows the ENSO forecast issued in April 2008 for a suite of
dynamical and statistical models. Note that at this time, most models predict an emergence from the weak La Niña phase into a near-neutral ENSO phase over the next year or two.

**Model Forecasts of ENSO from Jul 2008**

It is generally believed that ENSO is more difficult to predict during certain times of year. Since ENSO events usually develop between April and June and last until the following spring, forecasting events once in progress during those times is fairly simple. However, predicting which phase ENSO will develop into (between March and June) is much more difficult. In summary, ENSO forecasts issued between January through April are the least skillful. This is called the **spring barrier** in the Northern Hemisphere.

**References**


**Review Terms**

**Dynamical Models** –

**El Niño** -

**ENSO** -

**La Niña** -

**Multi-decadal oscillation** –

**Multivariate ENSO Index (MEI)** –

**Positive Feedback Loop** -

**Southern Oscillation** –
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