Overview of TEXMEX II

TEXMEX II Scientific Working Group*

February 13, 2004

1 Summary

Though tropical storm track forecasts have improved greatly in recent years, corresponding improvements in forecasts of intensity have not. Understanding tropical storm intensification and genesis (which is just a part of the intensification problem) are both of significant scientific interest and of great practical importance to the protection of life and property in coastal regions.

Recent advances in theoretical and modeling work suggest that the time is ripe to mount a major observational effort to understand the genesis and intensification of tropical storms. Furthermore, we now have observational tools which will allow us to make significant progress on these problems.

Tropical storms occur over all warm water tropical ocean basins. However, the eastern tropical Pacific experiences more tropical storms per unit area of ocean than any other region in the world. In particular, the region just to the south of Socorro Island (111° W, 19° N) suffers *more than four times* the incidence of tropical storms than the most affected region of the western Atlantic basin near the Yucatan peninsula (Rosengaus et al., 2002). The concentration of *cyclogenesis* is even more pronounced. The vast majority of these storms form and evolve within easy reach of modern airport facilities located along the Pacific coast of Mexico. Indeed, the east Pacific is a natural hurricane laboratory beyond compare.

The last 13 years have seen two major and a number of smaller meteorological field programs based in Mexico. The TEXMEX project (Acapulco, 1991) and EPIC2001 (Huatulco, 2001) demonstrated that large field operations in Mexico are feasible. Mexico is developing a cadre of investigators interested in tropical storms, so there are now real opportunities for collaborations between the US and Mexico in this area.

All of these factors suggest strongly that the most productive and cost-effective venue for studying tropical cyclogenesis and intensification is the eastern Pacific Ocean. TEXMEX II is our response to the demonstrated need to obtain a better understanding of nature's most destructive storms.

2 Background

Over the past 30-40 years, forecast errors for hurricane tracks have decreased by about 50%. This improvement is attributed largely to advances in observing capabilities and numerical weather prediction. However, such success has not been realized for intensity forecasts, with virtually no improvement seen in the past 20 years (DeMaria and Kaplan, 1997). The lack of improvement in intensity forecasting is the result of deficiencies in numerical models (e. g., resolution limitations

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and parameterization inadequacies), deficiencies in observations, and deficiencies in our basic understanding of intensity change. These deficiencies have significant consequences for risk management in the Atlantic and Gulf coastal regions of the United States and in the Gulf, Caribbean, and Pacific coasts of Mexico. Strong interest therefore exists in improving the accuracy of short-range (24-48 h) cyclone intensity forecasts and in extending such forecasts to longer periods, as evidenced by the fact that the National Hurricane Center began to make five-day track and intensity forecasts in 2003. The five-day forecast problem necessarily involves prediction of tropical cyclogenesis over oceans adjacent to populated areas. Thus, improvements in our understanding of tropical cyclone intensity change and tropical cyclogenesis (which is just the first step of intensification) would have significant practical benefits in improving intensity forecasts.

A 6-week field program was conducted in 1991 to document aspects of tropical cyclogenesis in the east Pacific. This project, called the Tropical Experiment in Mexico (TEXMEX), was carried out in Acapulco in July-August 1991. While important insights were gleaned from the TEXMEX field program, there were several areas of deficiency that can be identified. For example, observations suffered from a lack of accurate dropsondes, high-resolution airborne Doppler radars capable of sampling convective-scale updrafts and downdrafts, a lack of ocean measurements, and the inability to sample the entire depth of the troposphere. In addition, the use of a single aircraft compromised our ability to study simultaneously small-scale and disturbance-scale phenomena.

In spite of the shortcomings of TEXMEX, the project was able to study the genesis of four east Pacific tropical cyclones in a period of six weeks. This is a much higher rate than has been found to be possible in the Atlantic basin. The high availability of targets is due to the facts that (1) the east Pacific has the highest rate per unit area of tropical cyclogenesis in the world, and (2) the cyclogenesis region is easily accessible by research aircraft from the southern coast of Mexico.

In the 12 years since TEXMEX, we have developed a number of new observational tools pertinent to the study of tropical storms. Among these are high spatial resolution Doppler radars (NCAR's ELDORA), accurate dropsondes and expendable ocean probes, GPS navigation, and a wider variety and frequency of coverage of satellite measurements. In addition, new theoretical insights await to be tested against observations.

Several modern airports along the Mexican coast provide facilities usable by research aircraft. These facilities allow easy aircraft access to virtually all Pacific tropical storms east of about 115° W. In addition, there is growing scientific expertise on tropical storms in Mexico and an increasing awareness of the importance of this natural phenomenon to the economy, infrastructure, and public well being of this country. Thus, we can count on mutually advantageous collaborations with Mexican investigators in this project.

In September and October of 2001 a large field program called EPIC2001 (East Pacific Investigation of Climate) took place with the goal of better understanding the coupled ocean-atmosphere system in the tropical east Pacific (Raymond et al., 2004). One of the benefits of this project for tropical storm research was the development of significant additional insight into the climatology and variability of deep convection in the region. This knowledge will help in planning of future field programs there.

The convergence of new theoretical insights and improved observational tools indicates that the time is right for a major observational thrust on the issues of tropical cyclogenesis and intensification. The geography and climatology of the eastern Pacific region as well as the infrastructure and existing knowledge base available there strongly suggest that the east Pacific is the most appropriate venue for this project.

3 Scientific issues

Palmén and Newton (1969) summarized early work on the conditions needed to produce tropical cyclones. These include (a) sufficiently warm sea surface temperatures ($26° - 27°$ C); (b) sufficient distance from the equator ($> 5° - 8°$); and (c) weak vertical wind shear through the troposphere. To these conditions McBride and Zehr (1981) added the presence of a pre-existing tropical weather system, such as an easterly wave, in an environment with sufficiently strong cyclonic relative vorticity near the surface and sufficiently strong anticyclonic vorticity in the upper troposphere.

Many theories of tropical cyclogenesis focus on the manner in which cyclones can intensify (or not intensify) in isolation from other disturbances in the environment (Ooyama, 1969; Rotunno and Emanuel, 1987; Emanuel, 1989, 1995; Zehnder, 2001). Such mechanisms are typically axisymmetric in essence, and often depend on the interaction between the thermodynamic characteristics of the core environment and convection. This is an important idealization which may hold the key to cyclogenesis and intensification in certain circumstances, but it is an eventuality which is rarely realized. A variety of other factors can enter to tip the balance for or against development.

An additional set of theories concentrates on the asymmetric structure of developing tropical cyclones and pre-cyclone disturbances. For instance, Montgomery and Kallenbach (1997), Montgomery and Enagonio (1998), and Enagonio and Montgomery (2001) focus on the way in which asymmetric vorticity distributions with small-scale structure evolve into the monolithic core vortex of a mature cyclone. These calculations are pertinent to the interaction between the core vortex of a cyclone and vorticity produced by deep convection in rainbands or other disturbances in close proximity to the core.

The upscale growth process driven by the vortical hot towers is illustrated clearly in the animation of the formation of tropical cyclone Diana, simulated with the MM5 model using horizontal grid spacing of 3 km on the finest mesh. The Diana animation is shown at

`http://eliassen.atmos.colostate.edu/eric/3kmplots/vorticity_diana.html`

Sadler (1976, 1978) noted the importance of the proximity of a tropical upper tropospheric trough (TUTT) disturbance to the development and intensification of west Pacific typhoons, while Pfeffer (1956, 1958) and Challa and Pfeffer (1980) noted the importance of the convergence of eddy angular momentum flux to tropical cyclogenesis. Much attention has subsequently been paid to the role of upper level potential vorticity anomalies in the formation and intensification of tropical storms (e. g., Molinari and Vollaro, 1989; Molinari and Vollaro, 1990; Bosart and Bartolo, 1991; Molinari, Skubis, and Vollaro, 1995; etc.).

The essential physical principles behind upper tropospheric influences appear to be captured by the theoretical work of Montgomery and Farrell (1993). In this model the quasi-balanced lifting at upper levels due to the approach of a tropopause potential vorticity anomaly is transmitted to the lower troposphere in regions where pre-existing moist convection lowers the effective static stability of the troposphere to near-zero. This lifting causes low-level spinup, especially in regions with pre-existing positive relative vorticity at low levels.

A requirement in most theories of tropical cyclogenesis, including that of Montgomery and Farrell, is the reduction of the effective static stability to low values in the core of the incipient cyclone. The key element here is the suppression of convectively produced downdrafts (Emanuel, 1995; Raymond, López, and López, 1998). The elimination of planetary boundary layer (PBL) outflows produced by downdrafts allows the inflow of updraft air with non-zero absolute vorticity to spin up the low-level circulation. The development of this circulation is tied to the development of the warm core characteristic of tropical cyclones, and leads to the production of the intense surface heat fluxes needed to drive the cyclone heat engine.

It follows that those factors controlling the existence and intensity of downdrafts are crucial to cyclogenesis and cyclone intensification. This is evident in the analysis of three cases of cyclogenesis observed in TEXMEX. The precursor to hurricane Enrique had extensive downdrafts during the period in which it was observed, and its spinup was significantly delayed. On the other hand, downdrafts were weak in the precursor to Guillermo, and it intensified rapidly. Fefa started out rapidly intensifying, but its development was temporarily arrested when dry air penetrated to the core of the system.

In each of the above cases, intensification only occurred when the shear averaged over the storm core was weak. The exact mechanisms by which shear suppresses tropical storm development and intensification remain unclear. There are two main hypotheses, respectively dynamic and thermodynamic in character. In the former category, a sheared environment is capable of mechanically disrupting the cyclone vortex (Jones, 1995, 2000; Reasor and Montgomery, 2001). In the latter, shear is thought to suppress development by ventilating the core of the system with dry environmental air, thereby suppressing convective updrafts and promoting downdrafts (Simpson and Riehl, 1958; Gray, 1968; Raymond, López, and López, 1998). This appears to be what happened in the cases of Enrique and Fefa discussed above.

In addition to their sensitivity to ambient sea surface temperatures, hurricanes are now known to be sensitive to the heat content of the ocean mixed layer, and to respond to the upper ocean cooling induced by vertical mixing and up and downwelling processes (Chang and Anthes, 1978; Shay et al., 1992; Schade and Emanuel, 1999). However, in the presence of strong oceanic currents and fronts which can locally increase the heat content, this upper ocean cooling may be suppressed, resulting in explosive storm development, as in the case of Opal in the Gulf of Mexico (Shay et al., 2000), and more recently in hurricanes Isidore and Lili (2002). The importance of ocean heat content to storm intensity was demonstrated for a large sample of hurricanes by Mainelli et al. (2002).

As shown by Jacob et al. (2000), the hurricane-induced cooling of the sea surface appears to be produced by three mechanisms: (a) immediate entrainment of cool water below the ocean mixed layer induced directly by strong wind stress; (b) upwelling associated with the divergence of wind-induced currents in the ocean mixed layer; and (c) delayed mixing and entrainment associated with shear instability, where the shear is produced by the wind-induced currents. Evaporative cooling of the ocean surface appears to play a lesser role than ocean mixing and upwelling. Models such as those of Schade and Emanuel (1999) and Emanuel et al. (2004) only consider mechanism (a), which is expressible in a one-dimensional column. Emanuel et al. (2004) note that the other mechanisms need to be considered for slowly moving storms. This adds considerable complexity to the problem, as mechanisms (b) and (c) are inherently three-dimensional.

For stationary or slow-moving storms, upwelling processes along the track are a result of the net Ekman divergence of the ocean mixed layer current field. As the thermocline approaches the surface, steady-state geostrophic motions are generated. For storms moving more rapidly, the translating surface wind stress pattern forces oceanic mixed layer currents to rotate anticyclonically with time with a period of oscillation close to the local inertial period. These wind-forced, near-inertial currents also rotate anticyclonically with depth, creating significant current shear across the ocean mixed layer base that causes vertical mixing by lowering the Richardson number to below criticality (Price, 1981; Sanford et al., 1987; Shay et al., 1992; Jacob and Shay, 2003).

Both immediate and delayed cooling due to mixing of deep water through the base of the ocean mixed layer depend on two factors, the depth of the mixed layer and the static stability of water below the mixed layer, as characterized by the buoyancy frequency N there. Deeper mixed layers result in less sensitivity to upwelling and less turbulent energy at the base of the mixed layer, whereas larger buoyancy frequency requires larger values of shear for the production of shear instability and associated mixing. Hurricane Opal strengthened suddenly as it passed over a Gulf of Mexico warm core ring, at least partly as a response to reduced ocean cooling resulting from the much deeper ocean mixed layer in the ring (Shay et al., 2000).

The case of east Pacific hurricane Juliette (2001) may reflect the effects of variability in thermocline buoyancy frequency. This category 4 storm at first produced very little ocean cooling. However, as it moved to the northwest, it produced decreases in sea surface temperatures of up to 6° C in its wake. This may be due to a decrease in the thermocline buoyancy frequency from roughly 20 cph to about 8 cph along the trajectory of the storm (Shay and Jacob, 2004). The ocean response to Juliette is illustrated in the animation given in

`http://storm.rsmas.miami.edu/~cook/hurr/juliette.gif`

3.1 Uncertainties in our knowledge

Though we believe we have developed a general understanding of the factors important to tropical cyclogenesis and intensification, we have not reached the stage where this knowledge can be usefully incorporated into prediction schemes. Some of this uncertainty is surely due to lack of routine operational data pertinent to the problem. However, such data are expensive to obtain, and we need to pinpoint the kind of information needed to enhance our predictive skills. In conjunction with this, we need to sharpen our basic understanding of the cyclogenesis and intensification problems. In the following, we outline the areas in which we believe our knowledge to be lacking.

3.1.1 Reduction of downdrafts

The mechanisms by which convective downdrafts are reduced, thus allowing low-level spinup and cyclogenesis to occur, need to be sorted out. Humidification of the mid-troposphere and vertical homogenization of the moist entropy are obvious candidates for this (Bister and Emanuel, 1997; Raymond, López, and López, 1998), but downdrafts may also be dependent on other factors such as convective morphology. For instance, weak shear may help suppress downdrafts by forcing precipitation to fall through cloudy updrafts rather than surrounding clear air regions. Recent work suggests that convective cells in vorticity-rich cyclogenesis regions may rotate, resulting in a reduction of entrainment via inertial stability (Montgomery, Nicholls, Cram, and Saunders, 2004). In this case the suppression of downdrafts may be local to a particular convective system, which then goes on to dominate the cyclogenesis process. Cloud physical issues may even enter, via their effects on the ratio of convective to stratiform precipitation. Indeed, precipitation characteristics varied greatly in the east Pacific during EPIC2001, depending on whether convection was occurring in polluted or clean air (Darrel Baumgardner, personal communication, 2003).

3.1.2 Organization of potential vorticity

Recent work suggests that tropical storms don't spin up in a continuous, smooth manner, but develop in multiple steps as strong convective bursts converge air and concentrate vorticity (Montgomery and Enagonio, 1998; Enagonio and Montgomery, 2001; Davis and Bosart, 2001, 2002; Hendricks, Montgomery, and Davis, 2004). Numerical simulations suggest that these convective bursts produce cloud-scale or mesoscale vortices, which then merge and symmetrize to form the cyclone core. However, observations of this process are limited and fragmentary, and it is not clear that all tropical storms form in this fashion.

Once a tropical cyclone is formed, one must consider the interaction between the central vortex and the potential vorticity produced by convection in the rainbands. The vertical structure of the potential vorticity anomalies produced varies between convective and stratiform regions, and the interactions between these anomalies and the storm core are almost certainly too complex to be represented by simple barotropic models.

3.1.3 Storm suppression by shear

As noted above, suppression of tropical cyclogenesis and intensification by shear can take place by at least two different mechanisms, ingestion of dry air (thus promoting downdrafts in the core) and dynamic disruption of the vortex. Numerical calculations have had some success in reproducing observed tropical storm structure (e. g., Marks et al., 1992; Franklin et al., 1993) in the presence of shear (Bender, 1997; Frank and Ritchie, 1999, 2001). In particular, low level convergence is most intense on the downshear side of the storm and weakest on the upshear side. Convective cells tend to form on the downshear side and wrap around the eyewall of the storm, releasing most of their precipitation on the left side of the storm, facing down shear. Interestingly, east Pacific tropical storm Fefa (1991) exhibited this structure in a sheared environment during TEXMEX (Raymond, López, and López, 1998). Understanding the mechanisms operative in shear suppression of tropical storms is crucial to better genesis and intensity predictions.

3.1.4 Mechanisms of upper level forcing

The main problem here is that we lack upper tropospheric data with sufficiently fine space resolution and enough coverage in time to test existing theories. We need to see how the cyclone vortex and the associated convection responds in particular to the proximity of an upper level trough. This presumably combines the stimulative effects of quasigeostrophic lifting with the deleterious effects of shear. Which effect dominates under various circumstances needs to be determined. As noted above, there may be little chance to investigate this potential influence on tropical storms in the east Pacific, but this needs to be verified.

3.1.5 Ocean mixing and upwelling and their effects on storms

The physics of ocean mixing and upwelling due to strong winds is reasonably well understood from a theoretical perspective at this point. The main problem is in obtaining sufficient information about the state of the ocean over which tropical storms are likely to pass. Sea surface temperature is insufficient in this regard, as it does not reveal the depth of the mixed layer. TOPEX/Poseidon satellite altimetry data can help here, as an upward bulge in the ocean surface is accompanied by a downward bulge at the thermocline, and hence a thickening of the mixed layer. However, such data do not give us a direct measurement of the buoyancy frequency in the thermocline.

The feedback of the modified sea surface temperature distribution on the storm producing it is somewhat more uncertain. If, as Emanuel (2004) suggests, only the prompt mixing is important in this regard, then the problem is tractable. On the other hand, if the delayed response of the ocean surface due to induced currents is important, then the problem is likely to be much more difficult, due to the complicated dynamics of these currents.

3.2 Local factors in the east Pacific

Several factors peculiar to the east Pacific must be considered when planning an experiment there.

1. Established wisdom holds that that most east Pacific hurricanes arise from African easterly waves impinging on the region (e. g., Avila et al., 2003). Our experience in EPIC2001 is

that this connection frequently can be established only in retrospect and that perhaps more attention should be paid to the role of alternate types of pre-existing disturbances in east Pacific tropical cyclone formation. For instance, east Pacific storm Lorena (2001) almost certainly formed out of the intense convection associated with a Tehuantepec jet impinging on the warm waters of the Pacific. This jet resulted from the interaction of a Gulf of Mexico cold front with the topography surrounding the Isthmus of Tehuantepec (Zehnder and Raymond, 2002).

2. Even in the majority of cases in which an easterly wave is the proximate cause of an east Pacific tropical cyclone, interaction of the wave with the terrain of southern Mexico and Central America plays a significant role in cyclogenesis in the region (Zehnder, 1991; Mozer and Zehnder, 1996; Zehnder, Powell, and Ropp, 1999).

3. The phase of the Madden-Julian oscillation (MJO) in the east Pacific has been shown to play an important modulating role in tropical cyclogenesis in this region (Molinari and Vollaro, 2000; Maloney and Hartmann, 2000). The mechanisms active in this modulation are not completely clear, but may be via the direct effect of the MJO on east Pacific convection (Raymond et al., 2003), or via an indirect effect on the development or structure of smaller scale pre-existing disturbances such as easterly waves (Molinari et al., 1997; Maloney and Hartmann, 2001a,b).

4. Tropical cyclogenesis in the east Pacific takes place at low latitudes. The role of upper-level mid-latitude troughs is therefore probably less than in other regions. Similarly, the tropical east Pacific is far from the usual location of the tropical upper tropospheric trough (TUTT) which appears to be important to tropical cyclogenesis in the west Pacific (Sadler, 1976, 1978). However, we do not really have the observations to rule out the existence of all forms of upper level influence on tropical cyclogenesis in the region.

5. The east Pacific warm pool extends roughly from 6° N northward to the Mexican coast. North of equatorial upwelling and the cold tongue, and well removed from coastal upwelling and downwelling, the oceanic warm pool is most likely the result of local thermodynamic and mixing processes as well as wind stress curl. While the mean winds tend to be generally weak, strong wind bursts are induced by intensifying easterly waves and tropical depressions. The upper ocean is fairly complex in this regime as the thermocline shoals from west to east, which causes the ocean mixed layer to be relatively thin, but with very strong vertical gradients in the density structure (Raymond et al., 2004). Turbulent heat flux at the base of the ocean mixed layer is an important component to the heat balance due in part to mean current shear. Within the context of shear-instability (i. e. lowering the Richardson number to below criticality), the shallow thermohaline structure will inhibit extensive deepening of the layer during weak to moderate atmospheric conditions. A second important aspect of the mesoscale heat balance in the ocean mixed layer is horizontal advection by the currents. Given the north equatorial current and north equatorial counter current, the advective time scale in this regime, defined by the ratio of the scale of the current system and its representative velocity, is typically $1 - 2$ d. Thus, the upper ocean variability has a relatively short-time scale in this regime, and appears to be matched to transient atmospheric variability such as tropical cyclones and waves. The east Pacific contains many westward-moving warm core rings produced by episodic wind stresses in the Gulf of Tehuantepec and by the instability of Kelvin edge waves propagating along the coast, which deepen the ocean mixed layer over regions a few hundred kilometers in diameter (Zamudio et al., 2004). These rings may have a

significant effect on tropical storm intensification.

6. Due to the sea surface temperature distribution in the region, many east Pacific storms move over cold water shortly after reaching hurricane intensity. This provides an opportunity to examine the weakening and dissipation of tropical storms as their supply of energy from the sea surface is cut off.

7. Occasionally east Pacific storms make landfall on the coast of Mexico. As the Pacific coast is quite mountainous, this provides an opportunity to study the interaction of hurricanes with mountainous terrain.

3.3 Scientific questions

In light of the above comments, we propose the following set of questions:

1. *What conditions promote the suppression of downdrafts in cyclone precursors?* Obvious candidates are high relative humidity in the core and weak vertical gradients of moist entropy. Sensitivity of convective systems to shear and the possible suppression of entrainment by convective cell rotation need also to be explored, as do cloud physical issues. We especially need to explore the possiblity that downdraft suppression begins locally in isolated convective systems, which then spin up into the new cyclone core.

2. *How are asymmetric potential vorticity distributions symmetrized and incorporated into the cyclone core?* This question applies to both the pre-cyclone and mature cyclone cases. In the former, mesoscale convective systems in the pre-existing disturbance produce rather random patterns of potential vorticity. In mature cyclones, rainbands produce less random, but nevertheless asymmetric potential vorticity patterns.

3. *How does shear suppress tropical cyclogenesis and intensification?* Drying out of the core by ventilation is one obvious candidate. Disruption of the vortex is another. It will be difficult to separate these causes observationally, due to the possibility that ventilation always accompanies shear. Numerical modeling will be particularly important here.

4. *Under what conditions do storms suppress themselves due to self-induced upwelling of cold ocean water?* Answering this question requires first of all that both the prompt and delayed the changes in ocean structure due to the passage of a tropical cyclone be documented. Existing work suggests that the direct effects are more important to the cyclone in most cases. However, slowly moving storms may be more strongly affected by the delayed response of the ocean. The distribution of ocean disturbances such as warm core rings is important here.

5. *What is the role (if any) of upper level disturbances in east Pacific tropical cyclogenesis and intensification?* We suspect a minimal role in this region, but we need to either rule out this process or document its effects.

4 Observational requirements

To answer the above questions will require a series of observations of a number of tropical storms as they evolve. Measurements over the entire storm at mesoscale resolution, observations of the storm environment, and more detailed observations of the storm core and neighboring rainbands must be made. Furthermore, measurements must be made through the full depth of the troposphere.

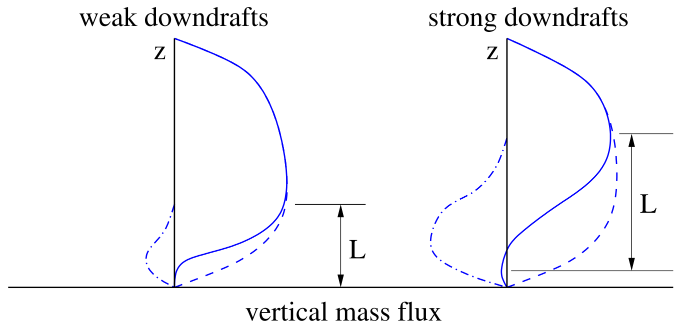

Figure 1: Sketch of the typical net vertical mass flux (solid lines) for a typical updraft profile (dashed) and different downdraft profiles (dash-dotted). The layer L over which net inflow occurs is typically less when downdrafts are weaker.

4.1 Measurement strategies

Not all of these questions need all possible measurements to be answered. What follows is a discussion of the measurement strategies for each question.

4.1.1 Downdraft dynamics

Raymond, López, and López (1998) established an indirect method of evaluating the strength of downdrafts. Mesoscale divergence is determined using standard Doppler radar techniques, from which the net vertical mass flux is determined. As illustrated in figure 1, updrafts often draw from roughly the lowest 3 km (Kingsmill and Houze, 1999; Mechem, Houze, and Chen, 2002). If downdrafts are weak, then the updraft inflow controls the net inflow. On the other hand, if downdrafts are strong, the net inflow feeds both the updraft and the downdraft, the latter of which draws from a much deeper layer – there may not even be net inflow at the surface.

The spinup of the inflow layer is mostly a balance between vortex stretching associated with the convergence, and frictional spindown. The thinner inflow layer results in more intense spinup for a given updraft mass flux. Thus, weaker downdrafts promote the low-level spinup of a cyclone.

Doppler radar has the disadvantage in this context of missing convergence in regions not containing rain. Also, sea clutter contaminates measurements at the very lowest levels. We propose to overcome these problems by deploying a sufficiently dense array of dropsondes over the region of interest. Between Doppler radar measurements and dropsondes, accurate measurements of mesoscale net vertical mass flux profiles should be obtainable.

Fine-scale observations of convection and the related kinematic fields are needed to determine individual updraft and downdraft structure. These measurements require the high spatial resolution of NCAR's ELDORA radar. Such measurements are important to determine the role of individual convective systems in the genesis of cyclones. In order to study the evolution of these systems, the measurements need to be repeated on a sub-hourly basis for several hours.

In addition to simply measuring the mass flux profiles, it is important to understand how these profiles are produced. Dropsondes deployed on a grid with $0.5° - 2°$ spacing over the full disturbance will allow us to determine the moisture field, and thus assess the importance of dry air in producing downdrafts. Doppler radar observations and wind profiles will allow us to assess the gross morphology of convective systems. Thermodynamic observations need to be made to well above the level of minimum equivalent potential temperature, i. e., up to 400 hPa or above, in order to get a complete picture of the moisture field. Wind fields above this level can be obtained by Doppler radar if clouds extend to sufficient height.

There is a conflict between obtaining sufficient spatial resolution and broad enough spatial coverage in designing the dropsonde grid pattern. For this reason it would be desirable to have a second dropsonde aircraft which deploys sondes on a larger scale surrounding the disturbance in order to characterize its environment. A high-altitude jet would be particularly desirable in this context, so as to obtain the environment through nearly the full depth of the troposphere. As patterns evolve rapidly in tropical storms, care must be taken to deploy dropsondes in a way that minimizes temporal aliasing in calculations of divergence.

4.1.2 Symmetrization of potential vorticity

Frequent maps of the horizontal velocity field of tropical storms and their precursors are needed as proposed in section 4.1.1 in order to calculate the evolution of the vorticity distribution in storms. Times for symmetrization presumably scale with the rotation time τ at the radius R of the vorticity anomaly,

$$\tau = 2\pi R/V,$$

where V is the tangential wind at that radius. This ranges from a day or more for a broad, weak tropical depression, to an hour or so for a hurricane. Thus, only in the weak storm precursor cases will the flight-to-flight time interval be small enough to resolve temporally this process. However, for strong storm cases, the time variation occurring within a flight should allow resolution of this process if repeated observations can be made. Simultaneous Doppler measurements from multiple aircraft in different parts of the storm may be helpful in this regard.

4.1.3 Shear suppression of cyclones

As noted previously, the two hypotheses for the weakening or suppression of tropical cyclones or their precursors are ingestion of dry air and mechanical disruption of the vortex. The strategy for investigating the first hypothesis is much the same as that described in section 4.1.1. This plan largely suffices for the second hypothesis as well, as the main measurement needed is the three-dimensional wind field. However, it is particularly important to obtain the wind field in the upper levels in this case, as some work suggests that the destruction of a tropical storm vortex by shear proceeds from the top down (Frank and Ritchie, 2001). Thus, obtaining dropsonde profiles from near the tropopause takes on a high priority.

In order to determine the effect of shear on a system, one must observe its evolution. Depending on the strength of the shear, a mature tropical storm responds to the shear in $6 - 36$ h according to Frank and Ritchie (2001). Given two radar and dropsonde-equipped aircraft, the minimum sustainable interval between observational missions is $12 - 14$ h. This observational interval should therefore be sufficient in all but the cases with strongest shear.

4.1.4 Production and effects of ocean cooling

The first task is to determine the state of the ocean previous to the passage of each tropical storm. The primary tools will be satellite observations of sea surface temperature and sea level altimetry. A three-dimensional pre-event survey would provide the necessary background to map the in situ profiler data to satellite-derived fields of SST and surface height anomaly, in order to assess the depth of isotherms and the ocean heat content. Periodic spot checks could then be performed as needed.

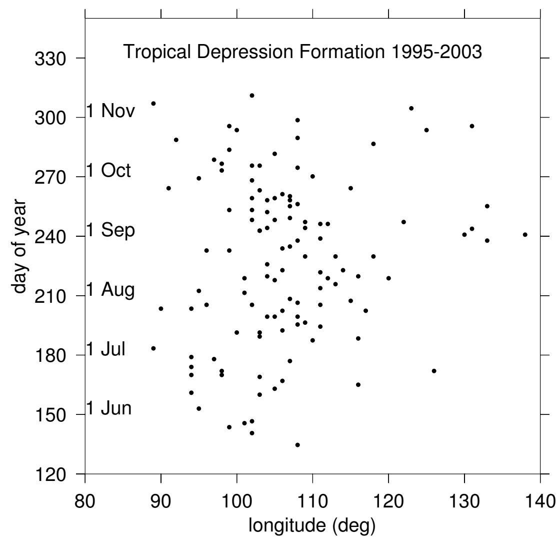

Figure 2: Day of year and longitude of formation of east Pacific tropical depressions which intensified into named tropical storms for the period 1995-2003.

The state of the storm can be measured using the same techniques described in sections 4.1.1-4.1.3. A factor worth considering is that expendable ocean probes generally need to be deployed with the aircraft unpressurized, which imposes a maximum flight level of about 700 hPa. A minimum level for deployment of about 810 hPa is also required for the probes to function properly.

4.1.5 Role of upper level disturbances

Addressing the effect of upper-level disturbances on tropical storms or their precursors is a difficult problem because the large-scale environment at upper levels must be characterized as well as the storm itself. We will have to depend largely on existing operational analysis schemes from NCEP and ECMWF to provide the big picture. Other than that the plans outlined in sections 4.1.1 and 4.1.3 should suffice, with the proviso that observations at repeated intervals (e. g., $12 - 14$ h) are crucial to reaching this goal.

4.2 East Pacific storm climatology

The National Hurricane Center tracks disturbances in the east Pacific, classifying them as tropical depressions, tropical cyclones, and hurricanes of various intensities. Historical results are available at

`http://www.nhc.noaa.gov/pastall.shtml?text`

We have used these data for the period 1995-2003 to learn something about the climatology of east Pacific cyclogenesis. Figure 2 shows a scatter plot of the longitude and day of the year at which all disturbances resulting in named tropical storms first became recognized as tropical depressions. Depressions are recognized to form west of about 90° W, with more forming west of 110° W late in the season.

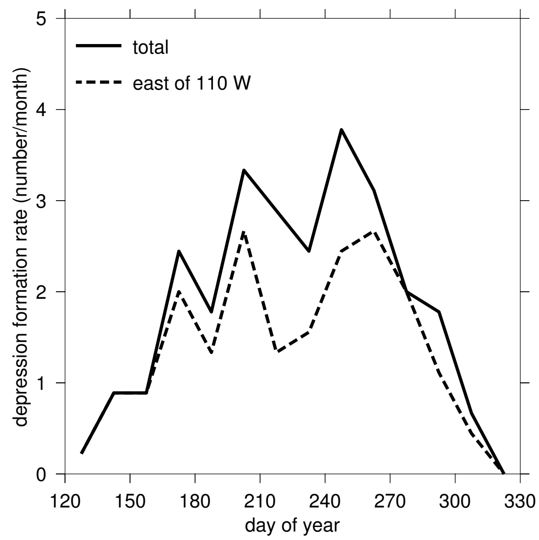

Figure 3: Frequency of tropical depression formation in the east Pacific as a function of day of the year. The solid line shows all cases while the dashed line shows all depressions forming east of 110° W.

Figure 3 shows that the peak formation rate occurs near day 250 (i. e., the beginning of September), but that for disturbances originating east of 110° W, the rate is approximately constant at two per month from the middle of June through the end of September.

Figure 4 shows the relative frequency of intervals between successive instances of tropical depression formation in the east Pacific. If depression formation were a random process, this curve would take the form of a decaying exponential. The shape of this curve shows that depression formation is far from random, with a peak interval of about 8 d and a long tail out to nearly 40 d. Depression formation thus occurs in bursts with variable intervals between bursts of up to 40 d. This is consistent with the analysis of Maloney and Hartmann (2000), which shows that tropical cyclogenesis occurs preferentially during the active phase of the MJO in the east Pacific.

Figure 4 poses a practical problem facing any study of east Pacific tropical cyclones. When the MJO is in its active phase, tropical storms are generated regularly with a mean interval between them of about a week. However, in the inactive phase, one may wait in the extreme case for a month or more for a cyclone to develop. Without attempts to predict the onset of an active MJO phase, one is therefore exposed to the real possibility that expensive observational facilities will sit idle for extensive periods. This is, of course, a fact of life for all tropical cyclone work, and it is less of a problem in the east Pacific than it is in the Atlantic basin. However, the cost of down time is greater, as U. S. facilities would necessarily be located away from their home base.

Two approaches to alleviate this problem are (a) to deploy facilities (i. e., aircraft) only when cyclogenesis is active in the east Pacific, and (b) to keep facilities in place and use them for alternate science projects during inactive periods. For instance, aircraft might be deployed to drop expendable ocean probes to obtain a background map of ocean properties during suppressed conditions, or to investigate non-developing convective clusters in order to determine why they don't develop. A combination of these two approaches may be appropriate for different platforms.

The first option would be most effective if an active MJO phase could be predicted by 48 h or more. MJO prediction is in its infancy (Waliser et al., 1999; Lo and Hendon, 2000; Wheeler and Weickmann, 2001). Matt Wheeler maintains an on-line real time prediction page for the MJO at

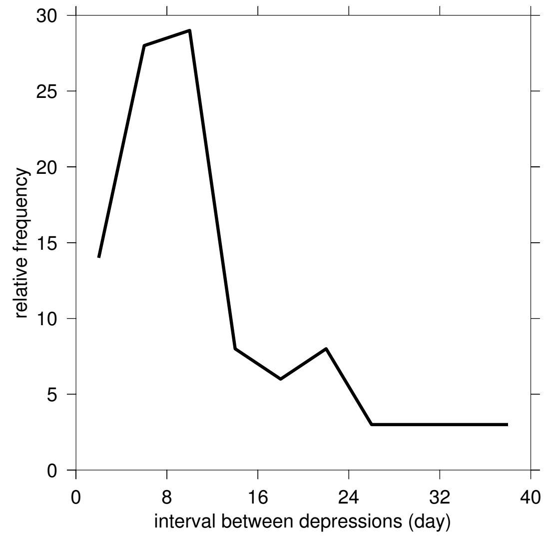

Figure 4: Relative frequency of intervals between successive instances of tropical depression formation in east Pacific.

`http://www.bom.gov.au/bmrc/clfor/cfstaff/matw/maproom/OLR_modes/index.htm`

which is based on statistical techniques using past and present outgoing longwave radiation (OLR) fields. The technique exhibits skill out to a couple of weeks, but our experience shows that it is not as useful in the east Pacific as it is in the west, due to the fact that the MJO tends to disappear in the OLR in the east Pacific. Reanalysis 850 hPa zonal winds may be more useful in the east Pacific for short-term predictions.

4.3 Complementary efforts

In addition to in situ observations, satellite data and numerical modeling are important to the success of this program.

4.3.1 Observational

Both satellite observations of the atmosphere and the ocean will be useful in real time to guide the project and in later analysis. Some of the most important satellite products are:

1. Geosynchronous visible and infrared satellite imagery has high time resolution, which helps to follow the development of disturbances of interest.

2. Quikscat surface winds are quite useful when they are available. Unfortunately, observations occur twice a day at best. These observations also deteriorate in heavy rain.

3. Passive microwave observations at various frequencies from SSM/I and TRMM satellites provide estimates of column-integrated vapor, cloud water, and precipitation. These are useful measurements for tracking the state of tropical disturbances. There are now sufficient satellites in orbit to provide several snapshots a day in many cases.

4. TOPEX/Poseidon sea surface altimetry will be extremely helpful for locating warm core rings in the ocean.

5. TRMM and other satellite-based measurements of sea surface temperature will be essential to the project.

4.3.2 Computational

Both atmospheric and oceanic numerical computations will play a role in the project.

1. High-resolution three-dimensional calculations of cyclogenesis and intensification cases will play an important role because in may cases only with this will we be able to infer causality. For instance, if a storm intensifies in a particular case, is it due to a decrease in the shear, motion over a deeper layer of warmer water, or favorable upper-level influences? Numerous groups worldwide are working on numerical models of tropical storms, including the NOAA group developing the HWRF model for operational use. TEXMEX II will provide detailed data against which modelers can test their computations. Once confidence is established in the fidelity of the numerical calculations, sensitivity tests can be used to evaluate the relative importance of various environmental influences. This knowledge can then be transferred to operational prediction models.

2. Recent progress has been made in the development of ocean models which now produce interesting predictions of ocean mixed layer evolution (Zamudio et al., 2004). Time scales for the normal evolution of the mixed layer are somewhat longer than the ≤ 5 d time scales being considered currently for hurricane prediction. Nevertheless, such results can provide necessary background for understanding the role of hurricanes in modifying the ocean mixed layer structure.

3. Coupled ocean-atmosphere modeling is still in a rudimentary state. However, simple coupled models such as that of Schade and Emanuel (1999) and Emanuel et al. (2004) show some skill in predicting tropical storm evolution. Such models need to be further explored and elaborated as needed.

5 Experiment design and facilities needed

Detailed experiment design at this stage is premature. However, some discussion of important factors in the design is appropriate. For the observational goals discussed above, we need the following measurements:

1. A grid of wind and thermodynamic profiles over the disturbance such as might be obtained an array of dropsondes. The spatial resolution can vary, with low resolution and broad extent in tropical storm precursor disturbances and higher resolution in tropical storms and hurricanes. The deployment pattern should be designed to produce as little time aliasing as possible, so that mean divergence and vorticity assessments are accurate. In addition, the pattern should be defined relative to the moving system, which generally means drifting the pattern to the west or northwest. Minimum drop altitude should be 400 hPa, with a desirable drop altitude of 200 hPa. An example of a possible grid pattern is shown in figure 5. This provides 0.5° resolution over a 3° square and can be executed in about 5 h, excluding ferry time, at a nominal aircraft speed of 4° h^{-1}. This pattern minimizes time aliasing for divergence and vorticity means centered on the storm. The pattern should be centered on the system of interest, and the initial part of the pattern serves to locate the center. This center will be better defined in tropical storms than in the precursor disturbances. Given the typical 10% failure rate of dropsondes, this pattern would require about 40 sondes per flight.

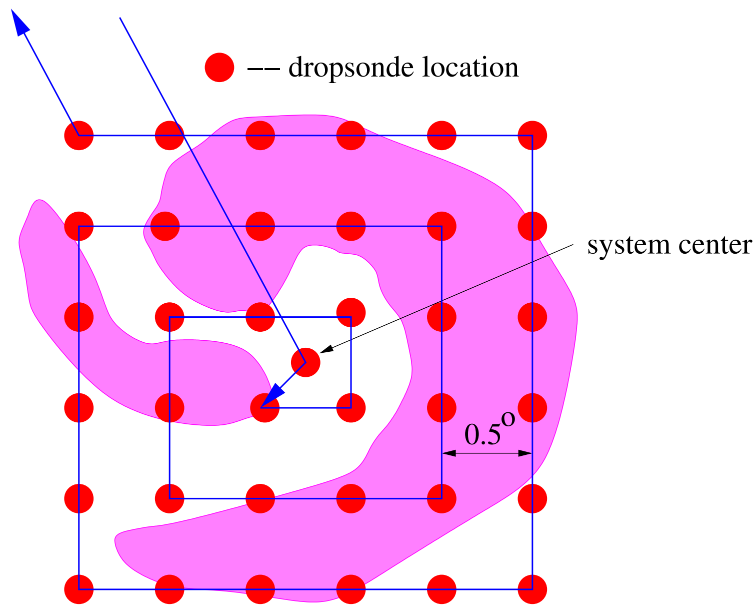

Figure 5: Suggested dropsonde grid pattern.

2. Doppler radar observations are needed to define the convective and precipitation structure of the system. The FAST scanning mode allows dual Doppler synthesis of winds in two swaths, one on each side of the aircraft, out to a maximum range of about 75 km with the NOAA P-3 radars. Doppler radar is also useful in inferring the winds in cloudy regions above the flight level of the aircraft. Surveillance radar observations out to a few hundred kilometers are important for determining the structure of the system of interest in real time and in locating a central starting point in the case of the pattern illustrated in figure 5.

3. Expendable ocean probe deployments are need in the stronger storm systems. These could be deployed on a grid similar to that shown in figure 5. Ocean probes generally need to be deployed between 6000 ft and 10000 ft, high enough for the probe to operate properly, but low enough for the deploying aircraft to operate unpressurized – a requirement for ocean probe use. Other deployments will be needed to survey ocean mixed layer conditions previous to tropical storm passage.

4. High space-time resolution Doppler radar observations (such as is supplied by NCAR's ELDORA radar) are needed to resolve details of the spatial structure of convection, e. g., to be able to evaluate separately the contributions of updrafts and downdrafts to the net convective mass flux, and to determine whether the convective system is rotating, as hypothesized by some investigators.

Inevitable conflicts exist between the procedures needed to attain various goals. There are two particularly important examples. Some goals need multi-aircraft missions to obtain near-simultaneous measurements in different regions of a system, while others need to use available aircraft sequentially to obtain the highest possible frequency of system snapshots. The other important conflict is the need to deploy dropsondes from high altitudes versus the inability to deploy expendable ocean probes from above 10000 ft. The proposed work is necessarily resource-intense, and careful planning will be needed to extract the most value from the resources available.

5.1 Potentially available facilities

Aircraft are the only in situ platforms capable of following a tropical storm as it develops. The following list indicates aircraft which (in the best of all possible worlds) could be made available to the project:

1. One or two NOAA P-3 aircraft. These come equipped with basic in situ measurements, surveillance radar, scanning Doppler radar, dropsonde and expendable ocean probe deployment capability. The maximum operating altitude is near 400 hPa.

2. NOAA G-IV jet. This has the primary role of deploying dropsondes. The maximum operating altitude is near 200 hPa.

3. NRL P-3 aircraft with ELDORA radar. This NSF-sponsored aircraft could be equipped to deploy dropsondes. Whether expendable ocean probes could be deployed is uncertain. This aircraft won't operate in hurricanes without guidance from a NOAA P-3 and will not penetrate eyewalls. Maximum operating altitude is near 400 hPa.

4. NASA DC-8 aircraft. This has extensive remote sensing equipment and can deploy dropsondes. The maximum operating altitude is near 200 hPa.

5. NASA ER-2 aircraft. This can deploy some dropsondes. The maximum operating altitude is in excess of 100 hPa.

6. Chartered jet aircraft for dropsonde deployment if no other high altitude aircraft are available.

NASA plans to use its two aircraft in CAMEX-5, a study of tropical cyclogenesis most likely based in San Jose, Costa Rica. They would be capable of operating as far west as 100° W. Generally speaking, they can therefore study cyclogenesis, but would not be able to reach the region where systems typically reach hurricane intensity.

5.2 Project scenarios

The minimum useful contingent of aircraft would be a single NOAA P-3 or the NRL P-3 for cyclogenesis studies and a NOAA P-3 and a high-altitude jet for hurricane intensification. In the former case the P-3 would deploy dropsondes from near 400 hPa and make Doppler radar measurements. Ocean probes are not needed, as systems near genesis are not expected to be strong enough to cause signficant upwelling and self-cooling. In the mature hurricane case the P-3 would fly near 700 hPa, deploying expendable ocean probes and possibly some dropsondes, while the jet would lay down a grid of dropsondes from near 200 hPa. The lack of a long-range surveillance radar in the NRL P-3 would put the burden of navigating the aircraft relative to the disturbance of interest on satellite observations. Fortunately, as we found in EPIC2001, the range of available observations has increased greatly since the days of TEXMEX.

The minimum project would preclude alternating aircraft to improve time sampling. Time resolution near 12 h rather than 24 h could be obtained with two NOAA P-3 aircraft, or with one NOAA P-3 and the NRL P-3 in the case of cyclogenesis.

In the likely case in which we had only one jet for launching dropsondes, upper level sondes probably could be deployed only every 24 h. This causes difficulty with any attempt to obtain snapshots of a storm every 12 h. Some compromises are possible here if ocean probes are deployed only when the jet is available for high altitude dropsonde duty. When it is not, the P-3 could fly high to obtain good dropsonde coverage.

The addition of NASA high altitude aircraft during the cyclogenesis phase would have the advantage that upper tropospheric flows and thermodynamics could be documented using dropsondes, thus obviating the need for another jet when this occurred. However, flight times and patterns consistent with the goals of CAMEX-5 would have to be worked out carefully with NASA.

The availability of two or three P-3s would allow two Doppler radar equipped aircraft to operate simultaneously under certain conditions. (As noted above, the NRL P-3 will not operate in hurricanes without guidance from a NOAA P-3.)

5.3 Operating location

There are a number of Mexican coastal airports from which research aircraft can operate. TEXMEX operated out of Acapulco while EPIC2001 used Huatulco. At various times NOAA/HRD has used airports further west.

Figure 2 shows that the highest density of tropical depression formation occurs in the range $100^\circ - 110^\circ$ W, but that depressions form as far east as 90° W. It is desirable (and possible, given our experience in TEXMEX) to identify potential cyclone precursors before the National Hurricane Center calls them tropical depressions.

Acapulco (99° W, 17° N) is located ideally to study the cyclogenesis process, as it is positioned just to the east of the maximum in depression genesis. Huatulco (96° W, 16° N) may be a bit far east, but this location would allow somewhat more overlap between our operational area and that of CAMEX-5. However, it must be kept in mind that Huatulco is not normally a 24 h airport, and this would put limitations on night operations.

For intensification studies, a site further west such as Manzanillo (106° W, 19° N) or Cabo San Lucas (110° W, 23° N) might be more desirable. Based at the latter location, studies could be made of deintensification due to movement of storms over cold water.

6 Project management

The project will be managed by its scientific working group, the membership of which is listed at the beginning of this document.

Cooperation with a number of different entities will be needed to make the project a success. These entities include:

1. NCAR/ATD (if we obtain the NRL P-3).
2. NOAA/HRD and the NOAA aviation facility. Two NOAA/HRD investigators are members of the Scientific Working Group.
3. NASA CAMEX-5 project. The potential here for working together on tropical cyclogenesis studies is high, and this connection needs to be cultivated. At least two members of the Scientific Working Group are also likely participants in CAMEX-5.
4. The NSF-sponsored hurricane rainband project of Shuyi Chen and Robert Houze (RAINEX).
5. The Mexican Government. Undertaking scientific work in Mexico is greatly facilitated by active collaborations with Mexican Investigators. Two Mexican scientists are members of the Scientific Working Group, one from Centro de Ciencias de la Atmósfera (CCA) of UNAM and one from Instituto Mexicano de Tecnología del Agua (IMTA). In addition, the chair of the Scientific Working Group spent a sabbatical year at CCA/UNAM in 2001-2002 while working on EPIC2001.

7 Data management

Plans for data management are not yet in place, but we will develop a scheme for making data available on the web.

8 Participants

At the moment this list includes only members of the Scientific Working Group. The list may expand in the future.

- Kerry Emanuel, Massachusetts Institute of Technology
- Tim Marchok, NOAA/NCEP, NOAA/GFDL
- Frank Marks, NOAA/HRD
- John Molinari, SUNY Albany
- Michael Montgomery, Colorado State University
- Ricardo Prieto, Instituto Mexicano de Tecnología del Agua
- Graciela Raga, Universidad Nacional Autónoma de México/CCA
- David Raymond, New Mexico Tech
- Rob Rogers, NOAA/HRD
- Nick Shay, University of Miami/RSMAS

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