Diurnal Analysis of Intensity Trends in Atlantic Tropical Cyclones

by

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Abstract

I postulate that a diurnal cycle may exist in observational variables related to tropical cyclone (TC) intensity. Prior studies document a significant diurnal signal in moist convection across tropical regions. Since convection becomes more pronounced in intense TCs, daily solar insolation possibly affects observed TC intensities. What remains unclear is if the diurnal signals in physical influences, or factors that modulate TC intensity over hourly timescales, are also prominent in observed TC intensity fields.

We apply various analytical techniques to two TC datasets and uncover a slight, yet detectable, diurnal trend in some calculated intensity fields. We first calculate 6-h maximum sustained surface wind (MSSW) tendencies using Atlantic TC best-track data from the National Oceanic and Atmospheric Administration (NOAA) National Hurricane Center (NHC) over years 1967-2011. In addition, we separate land tracks from warm-water tracks to analyze diurnal departures from the background states of these physical situations. We obtain a mid to late morning maximum in rapid intensification (RI) events over warm water. No discernable trend exists for landfalling TCs, even after using the decay model of Kaplan and DeMaria (1995) to find diurnal departures from mean decay rates. We also calculate theoretical TC indices using Atlantic TC dropsonde data from NOAA NHC over years 2002-2005 and 2011-2012. The indices, which measure physical influences on TC intensity, shift significantly during morning hours. This trend includes higher potential intensity (PI) and lower ventilation during late morning.

The diurnal signal in RI frequency and intensity indices follows prior statistics and two physical mechanisms. The signal’s greater PI coincides with more frequent occurrences of RI events, as confirmed statistically for Atlantic TCs by Kaplan and DeMaria (2003). Data noise likely obscures a possible diurnal signal in the negative MSSW tendencies analyzed in our study. Large-scale mechanisms that support our observed diurnal signal include enhanced radiatively driven low-level convergence and mid-level moisture during morning hours.
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Chapter 1

Introduction

Trends in tropical cyclone (TC) intensity on daily time scales comprise an important part of hurricane research today. Forecasting TC intensity even a few days in advance remains a challenge, since TCs can obtain and expel energy through various mechanisms. Important physical influences in a TC’s environment, which for example include sea-surface temperature (SST) and vertical wind shear, modulate intensity at hourly timescales. These influences can have major consequences for TCs making landfall in highly-populated regions, especially if the storm quickly intensifies. In response to this societal issue, numerous statistical and case studies have identified unique atmospheric environments that promote either TC intensification or decay. In our research, we focus on the diurnal aspects of TC intensity variability. Analysis of diurnal TC intensity trends is not widespread in current literature. TCs are a complex, large-scale phenomenon and research on their diurnal trends remains open for discussion and debate.

Our study investigates diurnal TC intensity variability using new methods of data selection. Instead of analyzing all TCs in general, we select data according to surrounding physical influences. We predict that diurnal trends can become more noticeable if analyzed under specific environmental conditions. But before we can select data for analysis, we must consider how TCs strengthen or decay in different situations. Most importantly, we should first discuss what physical influences control TC intensity and how those influences vary diurnally. Then we can decide how to
select intensity data, calculate their short-term trends, and examine them for any statistically significant diurnal signals.
Chapter 2

Significance of Short-term TC Intensity Trends

A part of TC research strives to reveal physical mechanisms underpinning short-term intensity trends. We use the adjective “short-term” as referring to hourly timescales and not exceeding one day. A combination of modeling, observed statistics, and case studies provides explanations of conditions, whether atmospheric or geographic, that control TC intensification and decay. Although many substantive ideas on short-term intensity trends exist in current literature, those regarding diurnal effects remain nebulous. Moist convection strongly responds to diurnal forcings and is the lifeblood of all TCs. Therefore, why are diurnal signals indiscernible in observed TC intensities? Diurnal effects should appear more pronounced under certain environmental conditions that surround a TC. Previous methods may have failed owing to improper constraints on data selection. This problem can give results that remain noisy and have no significant trend. We therefore synthesize past studies that examine short-term TC intensification rates to gather scientists’ current understanding of the phenomenon. We can use that information to devise an improved method of data selection and subsequent diurnal analysis.
2.1 Physical Influences

The heat-driven organization and intensification of TCs involves some key atmospheric conditions. TCs usually form and strengthen in favorable thermodynamic environments over the open sea. The disequilibrium between the sea-surface temperature and overlying air temperature is the source of TC intensification. This feature is an integral part of the wind-induced surface-heat exchange (WISHE) mechanism introduced in Emanuel (1986). Latent and sensible heat fluxes from the sea surface below contribute to both growth and maintenance of TCs, including areas lacking convective available potential energy (CAPE) (Emanuel, 1986). In addition, mesoscale convective systems (MCS) are significant precursors to TC development over warm ocean waters. In the Atlantic, the MCS usually forms within westward-propagating waves that originate in sub-Saharan Africa where the potential vorticity (PV) gradient changes sign. These waves usually require several days to evolve into tropical storms (DeMaria and Kaplan, 1994). The MCS, however, only strengthens into a TC under favorable atmospheric and oceanic conditions. Factors that contribute to genesis include a mixed-layer ocean temperature greater than $26^\circ \text{C}$ (e.g., see Palmén 1948; Gray 1968, 1978), weak vertical wind shear, a significant value of the Coriolis parameter $f$, low-level cyclonic relative vorticity, and a moist middle troposphere (e.g., see Lin 2007; Montgomery et al. 2012).

During and after genesis, high SST is arguably the most important factor in modulating intensity at daily timescales. This is because ocean waters hold most of the heat required to power TCs. The TC can absorb the ocean-stored heat once the surface wind increases. As defined in Lin (2007), the surface enthalpy flux, which contains the transfer of heat between ocean and atmosphere, is proportional to the surface wind magnitude. Then through the WISHE mechanism, updrafts carry the latent heat extracted from the ocean to form the warm-core vortex of a TC. Higher SSTs can substantially decrease a TC’s spin-up time. To illustrate this effect, we can
define an appropriate spin-up timescale as that in Emanuel (2003):

\[
\tau \approx \frac{H}{C_k V_{\text{max}}},
\]

(2.1)

where \( H \) is the atmospheric scale height, \( C_k \) is the enthalpy exchange coefficient, and \( V_{\text{max}} \) is the potential intensity (PI) of a TC (see Section 5.2.1 on theory regarding PI).\(^1\) DeMaria and Kaplan (1994) find that Atlantic tropical cyclones have maximum intensities at the 95th percentile ranging from \( \sim 40-60 \text{ m s}^{-1} \) over sea-surface temperatures (SSTs) of 26-30°C, with all TCs reaching \( \sim 55\% \) of the PI. Therefore, assuming that \( V_{\text{max}} = 50/0.55 \approx 91 \text{ m s}^{-1} \) in Equation 2.1, along with \( H = 10 \text{ km} \) and \( C_k = 10^{-3} \), \( \tau \approx 31 \text{ h} \) in areas conducive to TC genesis. This timescale is comparable to the estimate of \( \tau \approx 15 \text{ h} \) quoted in Emanuel (2003). Since higher SSTs produce higher PIs (e.g., see Emanuel 1987), the spin-up timescale decreases and short-term intensification becomes more probable. Therefore, TCs commonly spin-up in tens of hours following the formation of an MCS. This timescale appears in warm-water simulations of TC spin-up, such as that in Smith et al. (2009).

Environmental wind shear and mid-level moisture control short-term intensity through ventilation. TC ventilation, explored in earlier works such as Gray (1968) and Anthes (1982), involves the intrusion of dry air from the ambient wind shear. Ventilation hinders the growth and/or maintenance of a TC’s warm, moist core and therefore reduces intensity. These detrimental impacts can also occur over warm water and surpass the surface heat fluxes. The index of Tang and Emanuel (2012) assesses ventilation from an environmental sounding, with results depicting significantly lower (higher) ventilation indices for intensifying (decaying) TCs.\(^2\) Ventilation contributes to TC decay over short-term periods as well. The TC spin-up timescale of Equation 2.1 approximates the half-life of decaying TCs during spin-down if \( V_{\text{max}} \) and \( H \) are

---

\(^1\)First introduced in Emanuel (1989), this timescale is also the same as that for air to move vertically through a TC. The mathematical expression is essentially the TC’s depth scale divided by its vertical velocity scale.

\(^2\)Note that environmental soundings sample the atmosphere where no anomaly exists. These soundings, for example, do not include influences from TCs that may be in vicinity of the sampled location.
replaced with the storm-relative tangential velocity and tropopause height above the boundary layer, respectively. This spin-down timescale stems from theory presented in Eliassen and Lystad (1977) and has been confirmed with modeling of axisymmetric vortices done by Montgomery et al. (2001). Hence wind shear can ventilate TCs and significantly reduce their intensity over timescale $\tau$.

Interactions between ambient wind shear, ambient moisture, and underlying SST and their subsequent influence on TC intensity remain unclear. This lack of understanding is partly due to sampling issues. Emanuel et al. (2004) argues that uncertainty in vertical wind shear profiles contributes to uncertainty in intensity forecasting. Furthermore, they claim that PI, which depends on boundary-layer moisture, and upper-ocean thermal profiles influence intensity. More complete vertical atmospheric profiles could measure ventilation more accurately and improve short-term intensity predictions. However, limited resolution and data assimilation in numerical weather prediction (NWP) models follow as hurdles in achieving more accurate forecasts. Inaccurate intensity estimations from assimilating a portion of available infrared data and unknown cloud microphysics have major impacts on NWP model output (Park and Xu, Eds., 2009). Consequently, disagreements arise between modeled TCs and observed TCs under similar large-scale conditions. Cloud and moisture feedbacks within TCs is an ongoing motivation in research.

A TC finally dissipates when it loses its heat source. Both surface friction and the lack of warm water weaken TCs. For TCs that make landfall, surface heat fluxes are turned off and intensity rapidly decreases. The remnant circulation of a TC can survive over land at higher latitudes, but sometimes undergoes extratropical transition in the increasingly baroclinic environment. Kaplan and DeMaria (1995) provide empirical evidence that intensity in Atlantic TCs decays exponentially over land. Therefore the rate of weakening decreases over the period $\tau$ hours following landfall. Warm-core TCs may intensify over land at low latitudes but require a surface heat flux. Consistent with WISHE theory, the surface heat flux must be strong enough to maintain the TC’s warm core. This can happen for example in Australia, where hot soils with large heat conductivity evaporate rainfall fast enough
to power landfalling TCs temporarily (Emanuel et al., 2008). Without such fluxes, TCs lose their warm-core structure and tropical characteristics.

Sometimes TCs can significantly change strength over land from permanent surface features or large-scale dynamics. Modeled TCs in Shen et al. (2002) decay less rapidly if they traverse inland lakes that are deeper and larger in area. Despite their model’s agreement with the WISHE mechanism, observations of this phenomenon are quite limited for landfalling Atlantic TCs. Furthermore, the idealized nature of simulations is much simpler than the coverage and amount of land-surface moisture underneath observed TCs. More noticeable dynamical situations include increased PV advection into the TC vortex and forced moisture ascent from underlying topography. Higher cyclonic PV supports TC spin up and can maintain the circulation. Orographic forcings can enhance precipitation, but overall have differing effects on TC intensity and remain questionable.

2.2 General Statistical Analyses and Case Studies

Prior work on TC statistics and case studies focus on the significance and implications of short-term intensity changes. Since many physical interactions in TCs remain unknown, more observational analyses are necessary to distinguish which environments appear more conducive to intensification or decay. Current research is unable to physically explain the rapid intensification (RI) of TCs over daily time scales, whether these events occur over land or over water. This lack of explanation remains a problem for TC intensity forecasts, especially when a larger population is at risk for landfalling TCs. The increase in average Atlantic TC activity over past decades further compounds that risk as well (Emanuel, 2005). Insurance companies and legislators are concerned that RI events may increase with climate change, causing more loss of life and property damage (Balling Jr. and Cerveny, 2006). Unreliable forecasting of these events can be potentially catastrophic if they occur just
prior to landfall in a highly-populated region. Enhanced prediction of RI remains a top priority at NOAA NHC today (Rogers et al., 2013).

Several statistical studies have identified significant environmental conditions accompanying RI. Three definitions of RI include (1) an increase in maximum sustained winds of at least 15.4 m s$^{-1}$ (30 kt) in a 24-h period (e.g., see Kaplan and DeMaria 2003), (2) at least 5.14 m s$^{-1}$ (10 kt) in a 6-h period (e.g., see Balling Jr. and Cerveny 2006), and (3) at least 50 kt in a 24-h period (e.g., see Holliday and Thompson 1979). Using their RI definition, Kaplan and DeMaria (2003) claims that Atlantic TCs located over higher SSTs and higher relative humidity in the lower troposphere experience RI more frequently. A consequence of the higher SST is higher PI in the TC’s vicinity, giving favorable conditions for RI. More specifically, if the difference between the PI and the TC’s intensity are larger then the probability for RI is higher. Since Kaplan and DeMaria (2003) found their results significant over 24-h periods, diurnal effects cannot physically explain the results. Although Holliday and Thompson (1979) also use 24-h when defining RI, they recognize that the onset of RI happens more frequently at night for Pacific TCs. Only a few statistical studies analyze RI in Atlantic TCs over shorter intervals (e.g., see Balling Jr. and Cerveny 2006). One flaw inherent in some methods is the lack of isolating RI events according to prevailing atmospheric conditions. In future studies, this problem could be solved with model simulations of RI in specific environments. Whether or not the environmental aspects connected with RI vary diurnally remains open for further analysis.

The effects of RI are vivid in a few landfalling Atlantic TCs of the past two decades. Such events have been investigated in various case studies. For example, Hurricane Charley (2004) delivered a devastating punch to the Florida peninsula as it intensified from a maximum sustained wind of 95 kt to 130 kt in about eight hours prior to landfall (Pasch et al., 2011). Charley’s average intensification rate over the RI period was therefore 4.375 kt h$^{-1}$. Charley’s RI occurred during late morning and into afternoon, strengthening mostly from its interaction with an upper-level trough. Park et al. (2009) find that post-event model simulations of Charley require horizontal resolutions greater than 6 km to accurately reproduce the interaction and
anomalous RI. Another RI case is Hurricane Wilma (2005) which intensified from a tropical storm with winds of 60 kt to 130 kt in 24 h (Pasch et al., 2006). The RI of Wilma did not occur evenly over the 24-h period but more intensely as the eyewall became smaller. Wilma’s eyewall contraction was an extraordinary observation; Chen et al. (2011) suggests that detailed inner-core dynamics and storm size should be incorporated in operational forecasting of RI. Finally, Hurricane Opal (1995) was a striking case that demonstrated other key environmental conditions conducive to RI. Bosart et al. (2000) documented the case as a “jet-trough-hurricane interaction” that also passed over a warm-core eddy in the Gulf of Mexico. The same “explosive” characteristics are visible in some TCs that pass over the Gulf Stream, a warm ocean current that amplifies surface heat fluxes (e.g., see Kuo et al. 1991). Case studies on RI document each event’s prevailing atmospheric conditions extensively. These efforts try to evaluate known hurricane physics, and possibly uncover anomalous interactions that produce RI and require more statistical proof in future studies.

Another motivation in studying short-term TC intensity trends involves some rare cases of inland reintensification. When considering such cases, one must be cautious with those that occur near mountainous terrain. For example, Tropical Storm David in September 1979 reintensified near the Appalachian Mountains. Doswell III (2001) claims that diabatic heating on the storm’s east side tilted a nearby upper-level ridge and advected positive PV into the storm, located just east of a weak trough. Therefore both orographic and mid-latitude dynamical effects can contribute to TC intensification. Other TCs, however, reintensified over significantly different landscapes. One of the more remarkable cases is Tropical Storm Erin (2007), which made landfall in southeast Texas. The system weakened into a tropical depression as it traversed Texas, but then reintensified into a tropical storm over central Oklahoma during the hours of 00-06 UTC on August 19, 2007. Studies of this particular case have examined the effects of soil moisture content on the surface heat fluxes at the time of reintensification (Kellner et al., 2012; Evans et al., 2011). The strengthening of TCs over land require heat fluxes which may be aided by the moistening of the underlying soil and its subsequent increase in thermal diffusivity (Emanuel et al., 2008).
primary mechanism(s) contributing to overland reintensification remain debatable, yet important for understanding short-term trends in TC intensity and improving operational forecasts.

Detailed wind and moisture quantities can give more accurate statistics and improve short-term TC intensity forecasts. Although case studies provide details about anomalous TC intensification rates, extrapolating such information in real time is challenging. Forecast models predict Atlantic TC intensity and track with comparable skill at 12 h, but the intensity skill reduces to one half (one third) the accuracy of track skill by 36 h (72 h) (DeMaria et al., 2005). As noted, enhanced wind shear profiles could improve forecasts. Statistics on Atlantic TCs show that wind shear is crucial at modulating intensity at low latitudes (DeMaria, 1996). Incomplete shear profiles have led to calculations of bulk shear quantities, which can depend on as few as two horizontal wind field levels. One example measure, provided in DeMaria (1996) and used in Tang and Emanuel (2012) for estimating ventilation, is the difference between the 200 hPa and 850 hPa wind vectors. The discreteness inherent in bulk formulas also neglects smaller-scale processes that critically influence TC intensity. For example, Emanuel (2000) states that surface flux formulas ignore wave drag and sea spray, processes that influence air-sea interactions. Various bulk quantities used in TC statistics therefore may be measuring small-scale dynamics inaccurately. This problem unfortunately propagates as intensity errors in NWP model forecasts.

Statistics and case studies have verified many, but not all, physics underpinning short-term intensification rates. This research has been successful at confirming which large-scale physical influences are most significant in TC intensity changes. But sparse sampling of small-scale influences and the rarity of RI events prevent scientists from thoroughly explaining short-term trends. This problem has large societal implications in situations when NWP models cannot accurately predict intensity in landfalling TCs. Other less-threatening intensity changes, such as rapid decay, deserve equal attention. Unknown physical interactions may be occurring that reduce TC intensity in areas statistically conducive to intensification. Therefore, narrowing future statistics according to specific physical situations may uncover signals related to these mysteri-
ous interactions. Diurnal influences may be part of those interactions, and hence are the subject of our study.

2.3 Analyses and Hypotheses on Diurnal Effects

Given the significance of environmental factors modulating TC intensity, we question how daily insolation may affect TC intensity. Analysis of diurnal influences is a missing piece of the TC-physics puzzle. Given the rarity of events, case studies on RI lack investigation of diurnal influences. Statistics on short-term intensity trends have successfully uncovered a handful of large-scale environmental influences. Since solar radiation is mostly a large-scale factor in energy transfer, we shall undertake a statistical approach for our diurnal analysis. But first, we should review the diurnal signals evident in the tropics and decide how to extract them from TC intensity data.

The eyewall in organized TCs and outer rainbands is a product of diurnally-varying deep cumulus convection. Parcels ascending from the TC’s boundary layer transport high-entropy air, produced through WISHE, and subsequently feed ongoing convection. Early evidence from Gray and Jacobson, Jr. (1977) exhibits a morning maximum and an early-evening minimum in deep cumulus convection over tropical ocean regions. They conclude that tropospheric radiational cooling within an atmospheric disturbance (e.g. a TC) exhibits greater contrast with that of the surrounding, cloud-free environment between day and night. That contrast owes to direct absorption of solar radiation by water vapor during the day. Model simulations of tropical oceanic convection confirm the diurnal cycle, which peaks at predawn or early morning hours (Liu and Moncrieff, 1998; Yang and Slingo, 2001).

Since convective intensity in TC rainbands can vary diurnally, the same trends can occur in those bands’ emitted cloud-top radiation. Areas of active convection within a TC produce higher sustained surface wind speeds. Within the closed circulation of a TC, much of the high winds reach the surface through the downdrafts of convective cells. Infrared remote sensing is a popular method of measuring the brightness temperature of cloud tops, estimating their heights above sea level, and inferring
convective intensity. Infrared measurements of areal extent in TC cirrus canopies reveal a diurnal signal, but with varying minima and maxima across different ocean basins (Browner et al., 1977; Muramatsu, 1983; Lajoie and Butterworth, 1984). The study of Kossin (2002) reveals at least a semi-diurnal cycle in the cirrus canopies of individual TCs. Hence diurnal signals observed in ordinary tropical convection are also evident in infrared TC data.

Instability theory may explain the diurnal variations in TC convection. Hobgood (1986) argues that the cloud tops of TCs possess a diurnal cycle from net radiation fluxes. The fluxes cool (heat) the cloud tops during the night (day), hence steepening (flattening) the lapse rate in the troposphere and increasing (decreasing) the atmospheric instability. Higher upper-tropospheric instability usually means a greater probability of convective initiation or maintenance. Precipitation, which becomes heavier with intense convection, over the tropical oceans appears to follow the instability hypothesis (e.g., see Randall et al. 1991). These conditions further suggest a diurnal variation in CAPE. However, the notion of CAPE supporting the maintenance or spin-up of TCs, especially over the ocean, remains debatable in hurricane research. On the other hand, instability from air-sea disequilibrium over inland lakes varies diurnally in the model simulations of Shen et al. (2002). The same study observed the most pronounced diurnal signal for shallow inland lakes. This result suggests that diurnal variations may appear in TC intensity observations, especially for TCs making landfall over wet areas. TCs over ocean waters may exhibit these signals at smaller amplitude. The smaller amplitude would result from a larger subsurface-layer heat capacity (Shen et al., 2002).

Tropical precipitation studies have proposed mechanisms contributing to diurnal behavior. The dynamical explanation of Gray and Jacobson, Jr. (1977) states that radiatively-forced horizontal divergence modulates convection. This divergence although CAPE can be a good predictor of deep moist convection, it has relatively low daytime values over the tropical oceans compared to land areas. The vertical temperature profile in most tropical environments nearly follows the reversible moist adiabatic lapse rate ($\Gamma_m \sim 4-7 \, ^\circC \, km^{-1}$) of an air parcel originating in the boundary layer. Under these conditions, the atmosphere becomes neutral to conditional instability (Emanuel, 1994). The prevailing surface wind does not significantly affect convection that draws most of its energy from CAPE, unlike convection observed in TC intensification and the WISHE mechanism.
occurs in the cloud-free regions, where subsidence occurs at night as the upper troposphere cools. The subsiding air enhances low-level convergence in the convective areas and yields an early morning maximum. Another hypothesis states that nighttime radiational cooling increases overall relative humidity and reduces entrainment of dry air in convection (e.g., see Tao et al. 1996). Although these mechanisms are intended for explaining precipitation, proving their correlation with measured TC wind speeds requires more research. Furthermore, since the proposed mechanisms apply to large-scale observations, they should be tested against anomalies such as RI in TCs.

Given that TC cloud canopies and precipitation exhibit a diurnal cycle, we hypothesize that the mechanism(s) at play significantly influences TC maximum sustained surface wind (MSSW) speeds. Besides TCs possessing a nocturnal intensity maximum, their diurnal characteristics seem more prominent during periods of intensification (Tripoli, 2006). MSSW is an appropriate quantity for calculating intensification rates and drawing possible connections to RI. Cerveny and Balling Jr. (2005) use MSSW for performing a diurnal analysis of intensity and 12-h rates. Their analysis, however, only examines TCs over water. They do not isolate and examine the intensity data in other major physical situations, such as high SST and overland tracks. Cerveny and Balling Jr. (2005) and Balling Jr. and Cerveny (2006) find that TC intensification is higher during the daytime, supporting results in Lajoie and Butterworth (1984) and challenging the explanations of Browner et al. (1977) and Hobgood (1986). Despite their discrepancies, the diurnal cycle in TC intensity may rely on a final factor considered in our study. The compelling work of Elsner et al. (2010) has linked a greater response in tropopause temperature change in years with higher ultraviolet (UV) radiation. Note that solar UV oscillates every 11 years. Assuming that the Hobgood (1986) theory holds, diurnal variations in TC convection (and therefore MSSW) may become more noticeable during high-UV periods.

We test our hypothesis by isolating TC data according to specific physical situations. As we have discussed, physical factors strongly control short-term intensification rates. Hence the first part (Chapter 3) of our study examines overland TC tracks. Since TCs normally decay overland, we use the empirical model of Kaplan and De-
Maria (1995) to determine mean decay rates and analyze possible diurnal departures. We also isolate TC tracks located over warm water in the second part (Chapter 4) of our study. For these warm-water tracks, we apply other constraints to try and reduce noise that may obscure significant trends. In the third and last part (Chapter 5), we use warm-water dropsonde observations to calculate indices that assess the likelihood of intensification or decay. Thus, through a combination of landfall decay rates, warm-water intensification rates, and high-altitude dropsonde soundings, we attempt to find a diurnal signal in observed, short-term trends of TC intensity and its indices. We analyze our compiled results (Chapter 6) to identify any overlapping signals and attribute them to one or more physical mechanisms. We finally discuss and summarize the implications of our findings and outstanding problems for future investigations (Chapter 7).
Chapter 3

Landfall Decay Rates

In this chapter we focus on landfalling Atlantic TCs and examine characteristics of their short-term intensification rates. Since TCs usually decay in intensity over land, we cannot assume that positive rates are equally as likely as negative rates. Furthermore, we suspect that overland decay rates may appear noisy when plotted diurnally because land areas dramatically reduce TC intensity and increase the variability in decay rates. We therefore apply the Atlantic-TC decay model of Kaplan and De-María (1995) to calculate background (or basic-state) decay rates for three categories of landfalling TC intensities. Also, during that process, we document outcomes from the fitting procedure. In our results, we inspect the spatial and temporal distributions of intensification rates across the southeastern United States. We identify their important features and any inconsistencies that may affect diurnally-plotted data. The background decay rates represent a baseline for measuring possible diurnal anomalies in overland intensity. We search for those anomalies in the diurnal results and determine if they exhibit statistically significant trends.

3.1 TC Best-Track Dataset

The data used throughout our analyses follows the “best track” of every Atlantic TC from year 1967 to 2011. These data originate from the NOAA Tropical Prediction Center’s Best Track Reanalysis, documented in Jarvinen et al. (1984) and provided
through Emanuel (2013). The data includes estimates of each TC’s latitude and longitude at 0.1° resolution, 1-minute MSSW speed at 5-knot (or 5-kt) resolution, and minimum central pressure at 1-hPa resolution throughout the TC’s lifetime. These values are recorded at 6-h intervals, with observation times at 0000, 0600, 1200, and 1800 UTC. An overland status indicator also accompanies each observation, locating the TC’s center either over land or over water.

We only focus on TC data from years 1967-2011 due to issues regarding the instrumental record. Data since 1970 likely include every TC case that occurred over the Atlantic basin, along with estimates of each storm’s sustained surface wind speed (Emanuel, 2008). Hence most of the intensity data in our selected period follows more systematic measurement intervals. Prior to 1970, the data within the range 45-120 kt incorporates a wind speed adjustment stated in Emanuel (2013), considering findings in Landsea (1993). Although Atlantic TC intensity data after about 1958 appears more accurate according to Emanuel (2008), we analyze data from 1967 onward to make comparisons with results in Kaplan and DeMaria (1995). TCs outside the Atlantic basin are ignored given more significant observational gaps in other regions of the world.

In addition, we restrict our selected data to overland storm tracks within the southeastern United States and northeastern Mexico. In particular we limit the tracks to a land sector bounded at latitudes 24°N and 37°N and longitude 105°W. This region has relatively flat terrain, except for a small northeastern area that includes the Appalachian Mountains. Significant changes in elevation (e.g. mountains) can alter a TC’s movement, intensity, and primary circulation (Lin, 2007). Therefore we try to eliminate orographic influences using our sector bounds. Despite the fact that the NOAA National Hurricane Center stops issuing public advisories on storms that become extratropical, the data set includes points with extratropical status as well. Therefore, our sector omits the extratropical points that usually occur at higher latitudes and retains storms with tropical characteristics. Additionally, the sector mostly covers areas with systematic, ground-based observations and gives the best estimates of TC intensity. Similar reasons for choosing our bounds are stated in
3.2 Calculation of Local Solar Time (LST) and Top-of-Atmosphere Insolation

The local apparent sunrise, apparent sunset, and solar time are calculated for each observation as part of the diurnal analysis. We employ lower accuracy methods outlined in Meeus (1991) to calculate these values. The conversion from Universal Time Coordinated (UTC) to local solar time (LST) requires the equation of time, or difference between apparent and mean time, defined here in radians as:

\[
E = y \sin 2L_o - 2\gamma \sin M + 4\gamma y \sin M \cos 2L_o - \frac{1}{2} y^2 \sin 4L_o - \frac{5}{4} \gamma^2 \sin 2M, \quad (3.1)
\]

where \(\varepsilon\) is the ecliptic obliquity, \(y = \tan^2(\varepsilon/2)\), \(L_o\) the Sun’s mean longitude, \(\gamma\) the eccentricity of Earth’s orbit, and \(M\) the Sun’s mean anomaly. We set reasonable approximations for the mean ecliptic obliquity \(\varepsilon_0\), \(L_o\), \(\gamma\), and \(M\) to Equations (21.3), (27.2), (24.4), and (24.3) from Meeus (1991) respectively. We ultimately use these quantities to define the Sun’s declination \(\delta_o\) and hour angle \(H_o\) with equations (24.7) and (14.1) from Meeus (1991) respectively. The local solar time in hours is finally calculated as

\[
LST = UTC - \frac{4}{60}\lambda + E, \quad (3.2)
\]

where \(UTC\) and \(E\) are in hours, and \(\lambda\) is the longitude measured in degrees positive west of the Greenwich meridian.\(^1\) At exactly 12:00 LST, the sun’s angle above the horizon and solar insolation are both maximized. \(E\) is converted to hours by multiplying the result from Equation 3.1 by \(\frac{180}{15\pi}\) (Meeus, 1991). Whenever \(LST\) is negative, 24 h is added so that \(0 \leq LST < 24\). The apparent sunrise and sunset are respectively defined as \(AR = 12 - \frac{4}{60}H_o\) and \(AS = 12 + \frac{4}{60}H_o\) with \(H_o\) in degrees.

\(^1\)Note that LST commonly stands for Local Standard Time in most other literature. Unlike local solar time, Local Standard Time is a civil time. Therefore, 12:00 Local Standard Time does not necessarily represent solar noon.
Since we are concerned with TC data recorded over the last several decades, much longer astronomical timescales remain irrelevant here. For example, $\varepsilon \approx 23.4^\circ$ throughout our solar time estimations and the accuracy of $E$ is not compromised when applied to twentieth or twenty-first century data. Evaluations of the apparent sunrise, apparent sunset, and LST from our assumptions yielded errors of only $\sim \mathcal{O}(1-10 \text{ min})$ compared to results provided at NOAA ESRL (Accessed Aug. 2013). These errors reduced for positions closer to the equator. Given that best-track TC data is reported at 6-h intervals, we can safely neglect the inherent error in our calculation method.

We finally calculate the top-of-atmosphere solar insolation over LST intervals bounded by two observation times. We only focus on incoming top-of-atmosphere radiation given the complexities of radiative transfer at lower levels (e.g., see Liou 2002). Assuming that a LST interval has bounds $[T_1, T_2]$ h, the top-of-atmosphere insolation is defined as

$$Q = S \left( \frac{d_0}{d} \right)^2 \int_{T_1}^{T_2} \mathcal{F}(T) dT, \quad (3.3)$$

where

$$\mathcal{F} = \begin{cases} 0 & ; \quad 0 \leq T < AR \text{ or } AS < T < 24 \\ \cos \theta_0(T) & ; \quad AR \leq T \leq AS \end{cases}, \quad (3.4)$$

the solar constant $S = 1367$ W m$^{-2}$, the mean Earth-sun distance $d_0 = 1.496 \times 10^8$ km, $d$ is the Earth-sun distance for the current day, and $\theta_0$ is the solar zenith angle. Equation (24.5) of Meeus (1991) is used to find $d$. The resulting integration from Equation 3.3 provides the amount of solar energy received per unit area within the interval $[T_1, T_2]$ h. The zenith angle can be calculated using the hour angle, which changes throughout the day. From spherical geometry, the zenith angle can be expressed as

$$\cos \theta_0 = \sin \varphi \sin \delta_o + \cos \varphi \cos \delta_o \cos H_o, \quad (3.5)$$

where $\varphi$ is the latitude. Following the convention of Liou (2002), we assume that $\varphi$, $\delta_o$, and $d$ are constant when evaluating $Q$. Furthermore, we define the latter two constants at solar noon of the current day. Therefore the quantity $\cos \theta_0$ only varies...
with \( \cos H_0 \). Note that \( H_0 = 0^\circ \) at solar noon and increases (decreases) by 15° for every hour after (before) that time. An error of \( \sim \mathcal{O}(5 \text{ min}) \) exists for the sunrise and sunset time (i.e. when \( \theta_0 = 90^\circ \)) because \( Q \), as defined in Equation 3.3, is an approximate value. This error however remains significantly smaller than the LST intervals occurring between observations, which are normally 6 h.

### 3.3 Background Decay Rates from an Empirical Model

We begin our search for a diurnal signal in landfalling TC intensity by finding background decay rates for three intensity categories. We define these categories according to the TC’s best-track wind speed at landfall, with intensities of tropical storm, minor hurricane, and major hurricane. As with Kaplan and DeMaria (1995), we assume that the landfall wind speed \( V_0 \) is the first best-track point over land for any landfalling TC. We define each range of intensity using each landfalling TC’s \( V_0 \). The ranges are \( 30 \leq V_0 < 60 \text{ kt} \) for tropical storms, \( 60 \leq V_0 < 85 \text{ kt} \) for minor hurricanes, and \( V_0 \geq 85 \text{ kt} \) for major hurricanes.

Each TC’s landfall time is linearly interpolated based on distance, using the single 6-h segment that spans the coastline. Our procedure relies on two assumptions regarding each TC’s translation, that (1) the TC follows the interpolated track and (2) moves at constant velocity along that track over the six hours. The coastlines are graphed in the Global Self-Consistent, Hierarchical, High-resolution Shoreline Database (GSHHS) (see Wessel and Smith 1996). We use the GSHHS's low-resolution version with a linear interpolation between the polygon vertices that outline North America. This version of GSHHS appears more detailed than that used to determine the overland status of each TC. This issue arose when examining a few TC’s which were reported as over water from the best-track data, but were actually over land according to the GSHHS coastlines. We assumed the best-track landfall status values were correct despite the slight disagreement.
We choose landfalling TCs based on other constraints, in addition to those discussed in Section 3.1. We require that each TC remains over land at least 6 h, ignoring those that have only one best-track point over land. Also, each of those TC’s consecutive set of points must remain within our specified land sector. These requirements yielded 121 storms making landfall within the sector, with their tracks depicted in Figure 3-1.

We use the Inland Wind Decay Model (IWDM) of Kaplan and DeMaria (1995) to find the exponential decay rate $\alpha$ of each landfalling TC. The IWDM predicts the over land wind speed at time $t$ (h), where $t = 0$ at landfall, as

$$V(t) = V_b + (RV_0 - V_b)e^{-\alpha t} - C,$$  \hspace{1cm} (3.6)

where $V_b$ is the background wind speed, $R$ is a reduction factor for the initial wind speed, $C = m \ln(D/D_0) + b$, $D_0 = 1 \text{ km}$, $m = c_1 t(0-t)$, $b = d_1 t(t_0-t)$, $c_1 = 0.0109 \text{ kt}$
\( \frac{d_1}{d_2} = -0.0503 \text{ kt h}^{-2} \), and \( t_0 = 50 \text{ h} \). \( C \) is a correction term that accounts for the proximity of the TC to the coastline, where \( D \) is the distance (km) to the coastline. Throughout our study, we set \( V_b = 26.7 \text{ kt} \) and \( R = 0.9 \). The constants defined in Equation 3.6 are taken from Kaplan and DeMaria (1995) and are optimal for our region of interest. We find \( \alpha \) numerically using \( V(t) \) and \( D = D(t) \) as the dependent variables. From Equation 3.6 \( \alpha \) can be written as a theoretical linear slope:

\[
\alpha = \frac{-1}{t} \ln \left( \frac{V(t) - V_b + C}{RV_0 - V_b} \right). \tag{3.7}
\]

Substituting observed surface-wind velocities \( V_{obs}(t) \) for \( V(t) \) in Equation 3.7 and plotting the quantity \( \alpha t \) against \( t \) produces a line with positive slope for ideal cases. We apply a least-squares regression fit to this data and estimate its \( \alpha \). Observations with increasing or steady intensity over land shifts \( \alpha \) toward negative. Such changes in intensity can skew the mean exponential decay rate and ultimately the background decay rate. We ultimately wish to calculate a mean exponential decay rate for each of the three landfall-intensity categories.

We apply constraints to the fitting procedure to reduce the number of cases that may have anomalous intensity fluctuations following landfall. The first of these is that we only process wind records having \( V_{obs}(t) \geq 30 \text{ kt} \). Second, the coastline-proximity correction term \( C \) was only applied if it improved the correlation. Third, we did not alter the constants in Equation 3.6, in order to compare our results to those of Kaplan and DeMaria (1995). Fourth, we require the correlation coefficient \( r \geq 0.7 \) for three or more consecutive data points. Fifth, we ignored TCs that were already overland and subsequently entered the land sector.\(^2\) Sixth and finally, we restricted the maximum time after landfall \( t_f \leq 48 \text{ h} \). After applying our constraints, 51 of the 121 storms were successfully fitted to the IWDM. Of those 51 TCs, 18 benefited from the term \( C \) and therefore had higher correlations. Table 3.2 displays information for these 51 landfalling TCs, and Figure 3-2 gives their locations.

\(^2\)This condition applied to one TC making landfall in Mexico and moving into the land sector from the south.
<table>
<thead>
<tr>
<th>Name</th>
<th>Year</th>
<th>$\alpha$ (h$^{-1}$)</th>
<th>$r$</th>
<th>$V_0$ (kt)</th>
<th>$t_f$ (h)</th>
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<th>$V_0$ (kt)</th>
<th>$t_f$ (h)</th>
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Table 3.2: Above are all Atlantic TCs for years 1967-2011 making landfall in the land sector and yielding a successful fit to the IWDM. Note that TCs with two values in a column made two landfalls.
Figure 3-2: Above are IWDM-fitted tracks for the 51 landfalling TCs of Table 3.2. Green points are observed positions used in the fitting procedure to obtain $\alpha$ and $r$ values. Blue points show the remaining observed positions within the land sector. Red lines denote linearly-interpolated positions.

The significant reduction in total storms applied to the IWDM mostly results from our minimum wind constraint. If we accept TCs with $V_{\text{obs}}(t) \geq 25$ kt and lower $V_b$ to 22 kt, 71 storms follow the fitting requirements. Additionally, letting $V_{\text{obs}}(t) \geq 20$ kt and $V_b = 17$ kt yields 86 storms with IWDM fits. TCs that had steady or increasing intensity over land usually had low correlations when finding $\alpha$. Some of the eliminated TCs decayed rapidly just after landfall and had steady intensity afterwards. This situation applied, for example, to TCs Agnes (1972) and Gaston (2004). Lastly some TCs, such as Dennis (1981) and Isidore (1984), had steady intensity after landfall. This phenomenon mostly occurred over or slightly north to northeast of the Florida peninsula.

The mean exponential decay rate from all 51 storms is $\alpha_\mu = 0.097 \, \text{h}^{-1}$, in strong agreement with that of Kaplan and DeMaria (1995). The corresponding mean correlation and correlation squared are $r_\mu = 0.95$ and $r^2_\mu = 0.90$, respectively.\(^3\) As noted,

\(^3\)These values originate from 54 IWDM fits since 3 of the TCs made 2 separate landfalls.
lowering the minimum wind threshold increases the number of TCs used in the fitting procedure. Doing this however can have ramifications on the mean exponential decay rate. For example, if we let the minimum wind be 15 kt and \( V_b = 10 \) kt, 93 TCs give \( \alpha_\mu = 0.045 \; h^{-1} \). Despite this disagreement with Kaplan and DeMaria (1995), it follows the result in Emanuel (2000) which also fitted exponential curves out to various hours after landfall. Nevertheless, these fits were made beginning at the time of maximum intensity (which can be over water, preceding landfall) and assumed zero wind speeds for records that terminated at or before 96 h after time of maximum intensity. Therefore, as Emanuel (2000) states, a higher value for \( \alpha \) is probably due to how the curve’s initial intensity is assigned.

The background decay rate is evaluated for each intensity category using its corresponding mean exponential decay rate \( \alpha_\mu \). In other words, we calculate a different \( \alpha_\mu \) for each category since stronger TCs can decay more rapidly than weaker TCs over land. Hence the stronger TCs can have higher decay rates overall especially during times immediately following landfall. The mean exponential decay rates for tropical storm, minor hurricane, and major hurricane are \( \alpha_{\mu, TRS} = 0.083 \; h^{-1} \), \( \alpha_{\mu, M1H} = 0.113 \; h^{-1} \), and \( \alpha_{\mu, MAH} = 0.094 \; h^{-1} \) respectively. For verification purposes, we include a plot of modeled overland speeds versus observed speeds in Figure 3-3. Their correlation should be compared with Figure 5 from Kaplan and DeMaria (1995). The background decay rate is defined as the derivative of Equation 3.6, where we now let \( D = D(t) \) and use one of the three mean exponential decay rates depending on the intensity at landfall:

\[
\frac{dV(t; V_0)}{dt} = -\alpha_{\mu, V_0}(RV_0 - V_0)e^{-\alpha_{\mu, V_0}t} \]

\[
- \left[ c_1(t_0 - 2t)\ln \left( \frac{D(t)}{D_0} \right) + c_1 t(t_0 - t) \frac{1}{D(t)} \frac{dD(t)}{dt} \right] - d_1(t_0 - 2t). \quad (3.8)
\]

Note that the term \( \frac{dD(t)}{dt} \) requires calculations directly from the 6-h data. Hence we assume \( \frac{dD(t)}{dt} \approx \frac{D(t+3) - D(t-3)}{6} \) km h\(^{-1} \), with \( t \) in h.

We use the background decay rates to determine diurnal departures in observed decay rates. The observed decay rate is defined as the time rate of change in TC
Figure 3-3: The modeled velocities $V(t)$ of all landfalling tracks, including those not used to determine $\alpha_\mu$ values, are plotted against the observed velocities $V_{\text{obs}}(t)$. The values for $V(t)$ were calculated using the corresponding $\alpha_\mu$ for each track’s $V_0$. The plotted values are for $t \leq 48$ h and they are shifted by $V_0$ as well.

intensity over land for any 6-h period. From here on, since we frequently use this short-term intensity rate throughout our study, we will call it the “MSSW tendency.” We mathematically define the MSSW tendency as

$$V'_{\text{obs}}(t) = \frac{V_{\text{obs}}(t+3) - V_{\text{obs}}(t-3)}{6},$$

(3.9)

with $t$ in h. These tendencies are calculated for each best-track segment within the land sector and occur halfway between observation times. In other words, we assume these rates occur at 0300, 0900, 1500, and 2100 UTC. We also linearly interpolate their positions using the spatial coordinates from each 6-hr segment’s edges. We then apply methods from Section 3.2 to find the LST for each $V'_{\text{obs}}$. The diurnal departures are finally calculated as

$$\delta(T) = V'_{\text{obs}}(t) - \frac{dV(t)}{dt},$$

(3.10)

where $T$ is the LST of $t$. 

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3.4 Results

3.4.1 Spatial and Temporal Characteristics of Inland Decay

The spatial distribution of 6-h intensification rates across the southeastern United States is given in Figure 3-4. We only include “significant” rates to reduce clutter and ease inspection. These rates do not include the −5 kt and 0 kt 6-h tendencies, since the mean intensity rate is −0.693 kt h⁻¹ from the full distribution. Most of the high decay rates are located along the Gulf coast, west of Georgia and Florida. The highly negative tendencies are sparse, but remain near the coastline. The weakly negative tendencies, which occur more frequently, are located further inland. Hence the distribution in the Gulf-coast regions away from Florida follows an exponential decay pattern. Here the TCs decay most quickly during landfall, as observed by Schwerdt et al. (1979); Ho et al. (1987). This feature however does not apply near the Atlantic coastline of Georgia, South Carolina, and North Carolina. Here many rates are positive, which represents anomalous behavior for overland TCs. These rates occur more frequently closer to the land sector’s northeast corner. In addition, the area’s coastline is curved and can largely affect decay rates. Rogers and Davis (1993) have found that Atlantic TCs decay less (more) rapidly as they approach a concave (convex) coastline, relative to the ocean’s frame of reference. Although this behavior seems apparent along the sector’s Atlantic coast, the low frequency of overland intensification should not affect the diurnal statistics. This proposition also applies to the highly anomalous intensification over Oklahoma that occurred in Tropical Storm Erin (2007).

The temporal behavior of inland intensity decay is depicted in Figure 3-5. The plot visualizes exponential decay for intensities measured 48 h after landfall. Note that the graph’s exponential decay rate cannot be verified with Equation 3.7. This is because α, which is greater by ∼ 0.08 h⁻¹, represents decay toward $V_b$ and not zero. The logarithmic intensity changes are randomly distributed about the best-fit line. However the spread of this distribution increases with time after landfall; notice how the “cloud” of negative values becomes wider. Intensification-rate anomalies therefore
Figure 3-4: The 6-h MSSW tendencies (kt) in the land sector are plotted according to significant wind speed changes. The coloring interval follows the 5-kt data resolution. The −5 kt and 0 kt tendencies have been eliminated since their coverage is widespread. No +15 kt or +20 kt tendencies occurred during the period, whose colors have been whited out as well.

are more likely at later times. This does not necessarily mean, however, that diurnal anomalies would become more noticeable. As TCs move further inland after landfall, they encounter more hostile environments and can have large intensity variability. We can assume that the background decay rates are acceptable for measuring accurate diurnal anomalies closer to landfall.

### 3.4.2 Diurnal Arrangement of Decay Rates and Departures

All 6-h MSSW tendencies ($V'_{obs}$) are plotted diurnally in Figure 3-6. Specific insolation values are color-coded to distinguish 6-h intervals that overlap sunrise and sunset times. Additional information regarding the insolation, which does not exhibit any significant diurnal relationship with decay rates, is provided in Figure A-1 of the Appendix for reference. The data’s horizontal discreteness appears as 4 clusters of points; the land sector bounds and the likelihood of TC development during late summer and early autumn reduces the spread in LST. Based on the sector’s location,
Figure 3-5: The log change in MSSW above is measured relative to the velocity at landfall \( V_0 \). The best-fit line only encompasses data beginning at 6 h after landfall since \( V_0 \) occurs at the first overland best-track point. Only data for which \( t \leq 48 \) h are included. The line’s slope is \((-1.73 \pm 0.27) \times 10^{-2} \) h\(^{-1}\) at 95% confidence. This uncertainty is depicted by the dashed lines.
Table 3.3: Above are statistics for Figure 3-6. The number of points $N_t$, mean $\mu$, standard deviation $\sigma$, and standard error $\sigma/\sqrt{N_t}$ are given for each of the 4 data clusters.

<table>
<thead>
<tr>
<th>LST Interval (h)</th>
<th>$N_t$</th>
<th>$\mu$ (kt h$^{-1}$)</th>
<th>$\sigma$ (kt h$^{-1}$)</th>
<th>$\sigma/\sqrt{N_t}$ (kt h$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-6</td>
<td>175</td>
<td>-0.694</td>
<td>1.147</td>
<td>0.0867</td>
</tr>
<tr>
<td>6-12</td>
<td>178</td>
<td>-0.693</td>
<td>1.309</td>
<td>0.0981</td>
</tr>
<tr>
<td>12-18</td>
<td>173</td>
<td>-0.827</td>
<td>1.159</td>
<td>0.0881</td>
</tr>
<tr>
<td>18-24</td>
<td>164</td>
<td>-0.552</td>
<td>1.046</td>
<td>0.0816</td>
</tr>
</tbody>
</table>

from left to right the 4 clusters correspond to observations at 0900, 1500, 2100, and 0300 UTC respectively. In addition, the vertical discreteness originates from the 5-kt data resolution. We assume that the standard error is the uncertainty in each mean tendency, using the number of points $N_t$ at each observation time. The relevant statistics are summarized in Table 3.3. The noise across the clusters is widespread and span similar ranges in MSSW tendencies. Therefore, despite the relatively small errorbars, trends in mean tendencies are not statistically significant. We note however that the largest increase in mean MSSW tendency occurs between the intervals 12-18 LST and 18-24 LST.

The diurnal departures display anomalies from the background decay rates and are given in Figure 3-7. Their statistics are provided in Table 3.4 as well. The color-coded insolation follows the format of Figure 3-6, but no striking diurnal trend exists when graphing the insolation against departures (not shown). All the departures shown exceed a maximum model error, which varies with $t$ and the data’s resolution. This error gives appropriate bounds on the background decay rate. We assume this error is $\pm \alpha_{\mu,V_0}R(2.5)e^{-\alpha_{\mu,V_0}(t-3)}$ based on the data’s wind and temporal resolutions. We ignored propagating errors related to $C$, since most of TCs used in the IWDM did not need $C$ in Equation 3.6. 465 calculated departures remained outside of the maximum error and are displayed in Figure 3-7. However, we ignored one decay-rate departure of $+9.03$ kt h$^{-1}$ that required $\frac{dD}{dt}$ when evaluating $\frac{dV(t)}{dt}$; its 6-h segment spanned a large water inlet and made $D$ small and $\frac{dD}{dt}$ large. The somewhat bare region just below 0 kt h$^{-1}$ stems from the tendency resolution of $\frac{5}{6}$ kt h$^{-1}$. Mean
Figure 3-6: MSSW tendencies (kt h$^{-1}$) are plotted diurnally for the land sector. (a) shows the full plot, and (b) is a magnified section of (a). Mean tendencies and standard errors among all (negative) tendencies in each of the 4 clusters are shown in black (green). The red vertical bars enclose times of sunrise and sunset for all plotted points. Points with zero insolation are in yellow. The red points in clusters near sunrise/sunset possess nonzero insolation that is below the nonzero mean insolation of their respective cluster.
Table 3.4: Statistics for Figure 3-7, following the format of Table 3.3.

<table>
<thead>
<tr>
<th>LST Interval (h)</th>
<th>$N_t$</th>
<th>$\mu$ (kt h$^{-1}$)</th>
<th>$\sigma$ (kt h$^{-1}$)</th>
<th>$\sigma/\sqrt{N_t}$ (kt h$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-6</td>
<td>117</td>
<td>-0.475</td>
<td>0.842</td>
<td>0.0778</td>
</tr>
<tr>
<td>6-12</td>
<td>116</td>
<td>-0.492</td>
<td>1.042</td>
<td>0.0967</td>
</tr>
<tr>
<td>12-18</td>
<td>121</td>
<td>-0.594</td>
<td>0.859</td>
<td>0.0781</td>
</tr>
<tr>
<td>18-24</td>
<td>111</td>
<td>-0.327</td>
<td>0.704</td>
<td>0.0668</td>
</tr>
</tbody>
</table>

departures, calculated for each of the 4 clusters, along with their standard errors are provided in Figure 3-7. This process was also done for positive and negative departures separately.

The resulting three curves of Figure 3-7 display a subtle, but not statistically significant, diurnal signal. The highest occurrence of faster decay is in the 12-18 LST interval. Note how each curve possesses a minimum in that region. They also exhibit decay anomalies that increase during daytime, between intervals 6-12 LST and 12-18 LST. The black and green curves display the highest probability of intensification in the 18-24 LST interval. This behavior is similar to that of mean MSSW tendencies in Figure 3-6. Despite this agreement, the contradictory blue curve shows a maximum in the 6-12 LST interval. Trends between intervals 0-6 LST and 6-12 remain variable among the curves. The large vertical spread, in contrast to the small errorbars, echos that of Figure 3-6. A subtle trend, which appears in both Figures 3-6 and 3-7, occurs between intervals 12-18 LST and 18-24 LST. But we cannot confirm that a full diurnal trend exists in MSSW rates over land.
Figure 3-7: Same as Figure 3-6 but for departures from the background decay rates. Only departures that exceed the model error and have $t \leq 48$ h are plotted. The black and green points follow the description in Figure 3-6, but apply for the plotted departures. The magenta points are mean positive departures among the 4 clusters, with their standard errors shown as errorbars.
Chapter 4

Warm-Water Intensification Rates

Here we limit intensity data to warm-water regions. Our data selection represents a separate physical situation that encourages TC intensification, but on the basis of SST only. We do not use, for example, three-dimensional reanalysis data to apply other constraints on the large-scale conditions. We take a simpler approach and assume that warm SST sufficiently subsets intensity data to regions with high surface heat fluxes. Given the presence of those fluxes, we then search for possible diurnal trends in MSSW tendencies. The MSSW usually occurs in or close to the eyewall of organized TCs. Hence we are also isolating short-term intensification rates to those measured in/near TC eyewalls. Furthermore, we expect that positive and negative rates will be more equally distributed. This presumption is contrasted with results from the previous chapter, where overland rates are negatively biased.

4.1 Data

4.1.1 Sea-surface Temperature and TC Intensity

A diurnal analysis of TC intensity over warm ocean waters requires both sea-surface temperature (SST) and best-track data. Sea surface temperatures over the same period selected in Section 3.1 (1967-2011) are taken from the Hadley Centre Sea Ice and Sea Surface Temperature data set (HadISST1), archived at the Met Office Hadley
Centre and described in Rayner et al. (2003). The data set provides SST at monthly intervals and 1° grid resolution. For TC intensity, we use MSSW from the same best-track data as that described in Section 3.1. This time however we ignore all overland best-track points using each point’s overland status indicator.

### 4.1.2 Warm-water Track Selection

All best-track points that lie over Atlantic waters with SST $\geq 27^\circ$C are included for analysis. Note that we use the entire Atlantic basin and don’t apply any sector restrictions. This includes not specifying a northern latitude boundary, since our warm-water threshold remains at latitudes lower than those that commonly experience extra-tropical transitions. We calculate the SST for each best-track point first by finding its nearest, SST grid point. We then extract that grid point’s monthly-averaged SST from the previous, current, and proceeding months. Best-track points that lie directly between two or four SST points are considered to have SST values averaged over the neighboring SST points. We assume that the three SSTs occur at the middle of their respective months. The final, calculated SST is linearly-interpolated to the best-track point’s date.

The track selection also discards wind speeds measuring 40 kt and less. This limits the data to more organized TCs. Higher wind speeds occur more rarely and can fluctuate rapidly depending on surrounding physical influences. For example, a strong TC can quickly weaken when transitioning from warm to cold SSTs. However, we eliminate this common deterioration in TC intensity through the warm-water threshold. Hence the additional high-wind constraint should extract diurnal intensity changes more easily; the presence of high SST allows other environmental and/or radiational factors to modulate intensity and ultimately short-term rates.

### 4.1.3 Translation Speed Removal

We can manipulate MSSW data to reflect over-water winds by removing each TC’s translation speed. We estimate the translation speed at a best-track point using the
Figure 4-1: Map (a) contains all TC best tracks over warm water (SST ≥ 27°C). Observed, best-track points are in blue and linearly-interpolated positions are in red. Map (b) is an SST composite of months containing the tracks in (a), with a contour interval of 0.25°C. Note how the averaged 27°C bolded contour encloses most of the tracks.
mean distance traveled over the two neighboring 6-h intervals. Note that our method requires at least three consecutive best-track points to apply the speed correction at one point. We subtract the translation speed from the MSSW using an over-water wind equation from Schwerdt et al. (1979):

$$V_{\text{sym}} = V_{\text{obs}} - 1.17c^{0.63} \quad (4.1)$$

where $V_{\text{sym}}$ is the maximum symmetric wind speed and $c$ the TC translation speed. The coefficient of 1.17 was taken from Tang and Emanuel (2012). In the same sense as Equation 3.9, we denote the symmetric-wind tendency as $V'_{\text{sym}}$. In order to calculate the tendency, at least four consecutive points are necessary to yield at least two points with eliminated translation speeds. Despite this data requirement on each warm-water TC, the resulting spatial distribution of tracks (not shown) is very similar to that of Figure 4-1(a).

4.2 Results

4.2.1 Diurnal Arrangement of Data

Following the plot formats of Chapter 3, diurnal profiles of warm-water intensification rates are depicted in Figures 4-2 and 4-3. The former depicts normal distributions of MSSW tendencies in four clusters centered near zero. As previously noted in Section 3.4.2, the clustered behavior emanates from the points’ relatively small region on the Earth. No significant diurnal trend exists when considering all points given the large scatter and the minute shifts in mean value over different LST intervals. The same insignificant behavior results when selecting only positive or negative tendencies. The plot containing symmetric-wind tendencies conveys the same results as well, even though the computation of Equation 4.1 slightly reduces the discreteness of the MSSW. Given this reduction, other diurnal results that follow in Section 4.2.2 focus on the symmetric-wind data. Note that in Figure 4-3 the data constraints omitted 60 out of 341 total TCs over years 1967-2011 since those TCs had less than
Table 4.1: Statistical results of a two-sided $t$-test at 95% confidence are given for every combination of two 6-h LST intervals from the best-track MSSW (left-side) and symmetric wind (right-side) tendencies over warm-water. T/F indicates True/False for the null hypothesis.

Table 4.1 gives $t$-test statistics on the diurnally-partitioned distributions of MSSW tendencies and the symmetric-wind tendencies. These results can be compared with the black curves of Figures 4-2 and 4-3. The only significant difference exists between the 6-12 LST and 18-24 LST distributions. The trend between these intervals is decreasing intensification rates through the afternoon hours. This result however cannot be verified every 6 h and therefore lacks temporal details. Notice that a significant shift does not occur, for example, between the 6-12 LST and 12-18 LST intervals. Low $p$ values for $t$-tests (12-18, 18-24) LST and (0-6, 18-24) LST supports, but does not confirm, a minimum in rates during evening hours.

In order to consider more solar influences on intensity, we partition data from Figures 4-2 and 4-3 according to UV effects researched in Elsner et al. (2010). They claim that higher UV radiation, which is proportional to a higher Mg II index, warms the upper troposphere faster and leads to TC decay. Figure 2 of their study depicts a significant response in TC intensity and tropopause temperature when the Mg II index exceeds 0.274 (60th percentile). Therefore we now select our preexisting warm-water data according to years where Mg II > 0.274 by using their Figure 1, which
Figure 4-2: MSSW tendencies (kt h\(^{-1}\)) plotted diurnally for warm-water tracks. (a) shows the full plot, and (b) is a magnified section of (a). Mean tendencies among the LST intervals 0-6, 6-12, 12-18, and 18-24 are shown in black, with errorbars denoting their standard errors. Mean values for positive (negative) tendencies with their standard errors are in magenta (green). The red vertical bars enclose times of sunrise and sunset for all plotted points. Yellow points have zero insolation.
Figure 4-3: Same as Figure 4-2, but for symmetric-wind tendencies (kt h\(^{-1}\)).
<table>
<thead>
<tr>
<th>LST Intervals (h)</th>
<th>$p_{obs}$</th>
<th>$H_{n,obs}$</th>
<th>$p_{sym}$</th>
<th>$H_{n,sym}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-6, 6-12</td>
<td>.6671</td>
<td>T</td>
<td>.8631</td>
<td>T</td>
</tr>
<tr>
<td>6-12, 12-18</td>
<td>.9034</td>
<td>T</td>
<td>.7886</td>
<td>T</td>
</tr>
<tr>
<td>12-18, 18-24</td>
<td>.1398</td>
<td>T</td>
<td>.2652</td>
<td>T</td>
</tr>
<tr>
<td>0-6, 18-24</td>
<td>.2702</td>
<td>T</td>
<td>.1326</td>
<td>T</td>
</tr>
<tr>
<td>6-12, 18-24</td>
<td>.1249</td>
<td>T</td>
<td>.1736</td>
<td>T</td>
</tr>
<tr>
<td>0-6, 12-18</td>
<td>.7438</td>
<td>T</td>
<td>.6623</td>
<td>T</td>
</tr>
</tbody>
</table>

Table 4.2: Same as Table 4.1, except for years with high UV. The Mg II index from Elsner et al. (2010) measures the UV intensity. Three different periods are included in the statistics: 1978-1982, 1988-1991, and 2000-2002.

shows Mg II measurements from late 1978 to late 2007. We assume the high Mg II years are periods 1978-1982, 1988-1991, and 2000-2002. The resulting statistics of Table 4.2 however remain inconclusive that high UV affects MSSW tendencies diurnally. Although Elsner et al. (2010) find significant differences in intensity, the same may not apply to 6-h intensification rates.

4.2.2 Rates Histogram and Diurnal Departures

Full and diurnally-partitioned distributions of symmetric-wind tendencies are depicted in Figure 4-4. We use symmetric winds to reduce the discreteness of MSSW tendencies and produce statistics that are more representative of over-water surface wind speeds. We fit lines to the outer sides of the full distribution, as denoted by the blue points, but ignore the outliers. We assume these outliers are in intervals $V'_{sym} \leq -\frac{20}{3} \text{ kt h}^{-1}$ or $V'_{sym} \geq \frac{20}{3} \text{ kt h}^{-1}$. We concentrate on the line-fitted areas given that they are at least one standard deviation ($\sigma_{w,sym} = 1.121 \text{ kt h}^{-1}$) away from the mean ($\mu_{w,sym} = 0.323 \text{ kt h}^{-1}$) and include anomalous rapid decay and RI events. The same procedure to produce Figure 4-4 was done on MSSW tendencies with nearly identical results.\(^1\) A noteworthy feature of Figure 4-4(a) is the linear decay in $\log_{10}$-frequency on the distribution’s halves. The small spread in 95%-confidence slopes of

\(^1\)The full distribution of warm-water MSSW tendencies has mean $\mu_{w,obs} = 0.342 \text{ kt h}^{-1}$ and standard deviation $\sigma_{w,obs} = 1.088 \text{ kt h}^{-1}$. The distribution’s left (right) best-fit line, as that following the method to produce Figure 4-4(a), has slope $0.420 \pm 0.141 (-0.608 \pm 0.091) \text{ h kt}^{-1}$. The uncertainty is at the 95% confidence level.
Figure 4-4: The histogram in (a) is a full distribution of warm-water symmetric-wind tendencies, binned at a spacing of $\frac{5}{6}$ kt h$^{-1}$. The original distribution was normalized such that the 0 kt h$^{-1}$ frequency was unity. Its resulting log$_{10}$ distribution is depicted above. The orange lines are fitted to the blue points while ignoring the red points. The left (right) line’s slope is $0.492 \pm 0.037$ ($-0.595 \pm 0.068$) h kt$^{-1}$ at the 95% confidence level. These confidence ranges are traced in magenta. (b) depicts the same best-fit lines with log$_{10}$-frequency distributions containing the same data, but separated by LST intervals.
Figure 4-5: Diurnally-partitioned RI-rate departures in (a) are measured relative to the right-hand best-fit line of Figure 4-4. The mean of those departures are plotted diurnally in (b), with errorbars denoting their respective standard errors. The mean departures in (b) are placed at the centers of their LST intervals.

4-4(b) verifies this behavior. In addition, the left-hand slope has a smaller magnitude than the right-hand slope, exhibiting TC decay more frequently than TC intensification. Despite more occurrences of decay, the negative tendencies have random diurnal departures and no significant trend. Contrary to these results are positive tendencies, which exhibit slight diurnal shifts. Notice the generally positive departures for RI rates ($\frac{5}{3} \text{kt h}^{-1} \leq V_{sym}' \leq \frac{25}{6} \text{kt h}^{-1}$) over 6-12 LST and negative departures over 18-24 LST.

Figure 4-5 provides more detail on the diurnal shifts in RI rates. A significant peak (trough) in RI frequency occurs over 6-12 (18-24) LST. Between those intervals are insignificant departures and generally normal behavior. Less confidence exists with the diurnal minimum given its large errorbar in Figure 4-5(b). The diurnal maximum however appears consistent for three different RI rates ($\frac{5}{2}$, $\frac{10}{3}$, and $\frac{25}{6}$ kt h$^{-1}$). Therefore RI has its highest probability in the mid to late morning hours and lessens variably during other parts of the day.
Chapter 5

TC Intensity Indices from Dropsondes

Recall that convective instability is one proposed mechanism to explain diurnal variations in TC infrared data. CAPE is one such quantity that can assess the degree of instability but requires vertically-sampled atmospheric observations. Therefore we turn to dropsondes in this chapter, which provide observed, vertical atmospheric profiles of temperature and moisture. These data can be used to calculate CAPE and related quantities, giving a more detailed picture of the instability. But rather than analyzing CAPE alone, we explore if TC intensity indices possess diurnal trends. These indices include PI and ventilation, which (as explained in Section 5.2) both assess instability using the vertical data. A primary motivation for undertaking this endeavor is the diurnal signal in RI rates from results of the last chapter and statistics from Kaplan and DeMaria (2003). In that particular study, PI was higher during RI events.

In addition, similar to prior methods in this study, we select data based on significant factors modulating intensity. Once again we limit the data to warm-water areas. But we now consider physical influences located away from a TC’s eyewall. The MSSW is usually measured in the eyewall and does not convey information on the surrounding environment. The methods of this chapter ultimately isolate data in the outer edges of warm-water TCs. Since actual intensities are not measured
conventionally in these regions, we use observed dropsonde data to examine possible environmental factors promoting TC intensification or decay.

Dropsonde data from hurricane reconnaissance missions provide in situ observations, unlike infrared sensing, in the vertical at various times of day. This feature contrasts with most radiosonde (i.e. weather-balloon) data. For example, a day-night analysis has been done in Free et al. (2003) using radiosonde data at a few locations in the Caribbean. However, these data are only provided at 00 UTC and 12 UTC. Therefore generating a full diurnal picture of any possible trends would require data from stations at many different longitudes both in and outside the Atlantic basin. For our purposes, we are concentrating on diurnal trends in the Atlantic, where an abundance of dropsonde observations are taken over a broad LST range.

5.1 Dropsonde Data Set

The full dropsonde data set is provided through the NOAA Hurricane Research Division (HRD) and includes all observations made on NOAA reconnaissance aircraft. We select data from the Gulfstream-IV (N49RF) over years 2002-2005 and 2011-2012 with file extension .frd, which composes the set of “full-resolution data.” These files contain 3,221 soundings which have been operationally processed during missions using the Atmospheric Sounding Processing Environment (ASPEN) software of the National Center for Atmospheric Research (NCAR). Despite the real-time processing of the data, it does not meet thorough quality-control standards and NOAA HRD cautions that the “...data may occasionally be erroneous” (NOAA HRD, Accessed Aug. 2013). All the data include readings of pressure, temperature, relative humidity, zonal wind, and meridional wind for each dropsonde’s descent; a 1-s temporal resolution reflects the detailed vertical sampling. While importing the data for analysis, we required all points to have either “good” or linearly-interpolated values as labeled according to NOAA HRD (Accessed Aug. 2013). Each sounding usually exhibits such values for the relative humidity about 30 s after the first reading, most likely so that each dropsonde adjusts to the outside environment.
Our data constraints involve physical influences on PI and ventilation indices. We only select soundings located over warm water, with an SST ≥ 27°C. This requirement eliminates significant decreases in PI from cool SSTs, since Emanuel (1987) argues that the PI depends only on SST to a first approximation. We find the SST using the same data set and procedure described in Section 4.1. Another issue regarding data selection is that PI theory requires an environmental sounding. In other words, input soundings cannot be taken from anomalous conditions such as within TCs. Given that nearly all Gulfstream-IV dropsondes are launched in a vast array of locations and intensities within a TC, we apply a few surface restrictions to ignore soundings within or near the eye/eyewall. We require that the surface pressure be greater than 980 hPa and that the average wind speed within the lowest 20 m of the soundings is less than or equal to 20 kt. These requirements select soundings in locations closer to the TC’s surrounding environment. Furthermore, as another safeguard, the surface-pressure constraint removes erroneous dropsondes that stopped transmitting above the surface.

Lastly, each input sounding must have a top pressure less than 250 hPa. This requirement is sufficient for selecting soundings taken by the Gulfstream-IV. The aircraft’s soundings have a mean release (or top boundary) pressure of 179.9 hPa with standard deviation 11.6 hPa. Each sounding must extend high enough in the atmosphere, beyond the level of neutral buoyancy (LNB) to yield accurate CAPE and PI calculations. The LNB is located near the tropopause, which in the tropics has an average height around 150 hPa (K. Emanuel, personal communication). Most Gulfstream-IV soundings, however, are located below the tropopause and the LNB for parcels acquiring $CAPE_{RMW}^*$. To address this issue we append the top of each dropsonde profile using a composite temperature sounding from radiosonde stations in the Pre-Depression Investigation of Cloud-systems in the Tropics experiment, or PREDICT (2010) (e.g., see Montgomery et al. 2012). These radiosondes are contained in the project’s GTS Mandatory/Significant Level Sounding Data and

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1Dropsondes launched in TCs do not represent environmental soundings ideally, but their sampling over a large LST range and throughout the troposphere makes them attractive to the experimental aspects of our study.
provided through the NCAR Earth Observing Laboratory, sponsored by the National Science Foundation. The mean soundings averaged within ±2 h of 00 UTC and 12 UTC are given in Figure 5-1 for reference. The 00 UTC (12 UTC) composites apply to dropsondes from the 12-24 LST (0-12 LST) interval. The appending process begins with the temperature point on the composite that is closest to, and below the top of a dropsonde’s profile. The difference in temperature between this point and that of the dropsonde at the same pressure is used to shift the composite profile, so that the dropsonde profile is linked to the lower stratosphere.

5.2 Calculation of TC Intensity Indices

5.2.1 Potential Intensity (PI)

A quantity for diagnosing favorable environmental conditions on intensity is the potential intensity (PI) introduced in Emanuel (1986, 1988) and subsequently revised in Bister and Emanuel (1998, 2002). The PI of a storm depends on a TC’s specific energy balance. This balance consists of heat input from sea-surface evaporation and dissipative heating of a certain thermodynamic efficiency, and output from mechanical, boundary-layer dissipation (Bister and Emanuel, 1998). The PI is a theoretical upper bound on TC intensity, achievable at the radius of maximum wind (RMW) and only if the TC operates as a Carnot heat engine (Emanuel, 1986). As defined in Bister and Emanuel (2002), the PI is

$$V_{max}^2 = \frac{T_s C_k}{T_0 C_D} [CAPE_{RMW}^* - CAPE_{RMW}], \quad (5.1)$$

where $T_s$ is the SST (in K), $T_0$ is the outflow temperature (in K) of a saturated air parcel at the RMW, $C_k / C_D$ is the ratio of the enthalpy exchange coefficient to the drag coefficient, $CAPE_{RMW}^*$ ($CAPE_{RMW}$) is the CAPE of a saturated sea-level (ambient boundary-layer) air parcel at the RMW. Note how Equation 5.1 demonstrates that high SST increases PI and promotes TC genesis/intensification. The PI’s CAPE quantities are calculated relative to the environmental sounding, which is in this case
Figure 5-1: The composite vertical temperature profiles are computed from radiosonde stations across the Caribbean region during PREDICT (2010). The red (black) profile are averaged from data recorded up to two hours before or after 00 UTC (12 UTC), during the period of Aug 15 to Sep 30. The stations used to compute the profiles are as follows: Barbados (BDI); Curacao (ACC); Le Raizet, Guadeloupe (FFR); St. Maarten (ACM); Santo Domingo, Dominican Republic (SDQ); Kingston, Jamaica (KJP); and Nassau, Bahamas (YNN). The background tephigram (thin dashed lines) contains isotherms (red), saturation mixing ratio lines (blue), dry adiabats (black), and pseudoadiabats (green).
each dropsonde profile.

The CAPE calculations rely on the lifted parcel’s vertical buoyancy profile. We assume each lifted parcel undergoes pseudoadiabatic displacement. Although reversible ascent minimizes the LNB height, which can be useful in keeping the LNB below the dropsonde release level, the pseudoadiabatic assumption keeps calculations consistent with those for entropy and ventilation (described in Section 5.2.2). Therefore the buoyancy is proportional to the difference in virtual temperature between the parcel and its environment (Emanuel, 1994). The top, appended portion of each sounding assures that all lifted parcels reach the LNB and have negative buoyancy at the top. This method eliminates improper estimates of the environmental CAPE, or $CAPE_{env}$, as well as the CAPE values in Equation 5.1. Since virtual temperature varies with moisture content, the PI calculation also requires vertical profiles of the vapor mixing ratio

$$r_v = \frac{\epsilon e}{p - e},$$

(5.2)

where $\epsilon$ is the ratio of the dry gas constant to the vapor gas constant, normally approximated as $\epsilon \approx 0.622$, $e$ is the vapor pressure, and $p$ is the total pressure. $e$ is found from the data’s relative humidity (%) which is defined as $H = \frac{e}{e^*} \times 100$, where $e^*$ is the saturation vapor pressure. We finally solve for $e$ by approximating $e^*$ using the formula of Bolton (1980):

$$e^* = 6.112 \exp \left[ \frac{17.67T}{T + 243.5} \right],$$

(5.3)

with $T$ specified in degrees Celsius. This particular approximation is suitable for our data, for $e^*$ is accurate to $\pm 0.3\%$ for the range $-35^\circ C \leq T \leq 35^\circ C$ (Emanuel, 1994). The boundary-layer mixing ratio profile has the greatest impact on PI as opposed to

$^{2}$The LNB height is lowered since reversible processes include condensate loading and heating. The condensate makes the parcel more dense than if it were lifted pseudoadiabatically, especially at pressures greater than 150 hPa (Emanuel, 1994). Hence the parcel’s buoyancy is lower overall from the condensate, despite that its moist adiabatic lapse rate decreases from its nonzero liquid-water mixing ratio (e.g., see Equation 4.7.3 of Emanuel 1994).

$^{3}$The top-level buoyancy increases when performing the CAPE calculations in this order: $CAPE_{env}, CAPE_{RMW}, CAPE^*_{RMW}$. 

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that of the mid and upper troposphere. Therefore, we set \( r_v = 0 \) above the 400-hPa level on each sounding so that \( r_v \) accompanies \( T \) at every pressure level (including the appended portion). We perform all PI calculations using computer code provided at ftp://texmex.mit.edu/pub/emanuel/TCMAX/. To summarize, the PI calculation ultimately uses the vertical \( T \) and \( r_v \) profiles to find \( CAPE_{env} \), \( CAPE_{RMW} \), and \( CAPE^*_{RMW} \). \( CAPE_{env} \) serves to estimate the PI in terms of minimum central pressure, which is less than the observed surface pressure on all dropsondes. An iterative procedure solves for the latter two CAPE values, yielding the final PI estimate upon convergence.

5.2.2 Ventilation Index

We assess the degree of TC ventilation using the index of Tang and Emanuel (2012). Their ventilation index is

\[
\Lambda = \frac{u_{\text{shear}} \chi_m}{V_{\text{max}}},
\]

where \( u_{\text{shear}} = |u_{850} - u_{200}| \) is the large-scale bulk shear between 850 hPa and 200 hPa, and \( \chi_m \) is the mid-level entropy deficit. Since the dropsonde data possesses high vertical resolution, we assume that \( u_{850} \) (\( u_{200} \)) is the averaged wind field between the 825-hPa and 875-hPa (225-hPa and 250-hPa) levels. Note that the latter quantity is evaluated at lower heights due to variability in the dropsonde release level. The entropy deficit, as defined in Tang and Emanuel (2012), is

\[
\chi_m = \frac{s_{m}^* - s_m}{s_{\text{SST}}^* - s_b},
\]

where the various moist entropy quantities \( s_{m}^* \), \( s_m \), \( s_{\text{SST}}^* \), and \( s_b \) are respectively saturated at 600 hPa, ambient at 600 hPa, saturated at SST at the RMW, and ambient in the boundary layer at the RMW.\(^4\) We also follow their use of the pseudoadiabatic

\(^4\)As suggested in Tang and Emanuel (2012), the quantity \( s_{\text{SST}}^* - s_b \) is evaluated at the RMW. Here we assume the total pressure at the RMW is \( p_{RMW} = p_s \exp \left[ -\frac{1}{c_p T_s} \left( \frac{V_{\text{max}}^2}{2} + CAPE_{RMW} \right) \right] \), where \( p_s \) is the ambient sea-level pressure and \( c_p \) is the specific heat of dry air at constant pressure. This estimate is taken from Bister and Emanuel (2002) and assumes that the wind field is in cyclostrophic balance at the RMW.
moist entropy:

\[ s = c_p \ln(T) - R_d \ln(p_d) + \frac{L_v r_v}{T} - R_v r_v \ln(H), \]  

(5.6)

where \( c_p \) is the specific heat of dry air at constant pressure, \( R_d \) is the dry gas constant, \( p_d \) is the partial pressure of dry air, \( L_v = 2.555 \times 10^6 \) J kg\(^{-1}\) is the (constant) latent heat of vaporization, \( R_v \) is the vapor gas constant, and \( T \) is in K. As with \( u_{shear} \), we evaluate the 600-hPa moist entropy with averaged values of \( r_v, p, \) and \( T \) between the 550-hPa and 650-hPa levels. \( s_b \) is calculated assuming its \( T \) and \( r_v \) are from the lowest level in the sounding, the same level at which parcels begin ascending in the PI algorithm. The final moist entropy quantities yield an effective \( \chi_m \) that we use to calculate \( \Lambda \).

### 5.3 Diurnal Results

PI and ventilation estimates from dropsonde profiles are plotted diurnally in Figure 5-2. We encountered some anomalous values of \( \chi_m \) during calculations. As a result, we only display results that have \( 0 \leq \chi_m \leq 1.5 \) and thereby ignore outliers. This restriction on \( \chi_m \) reduced total soundings by 240. Along with other data restrictions discussed in Section 5.1, 1,376 dropsondes profiles remained with PI and ventilation estimates. In addition, temperatures at the tops of these soundings were 1.21°C warmer on average compared to the PREDICT (2010) composite, with a standard deviation of 1.34°C. A majority (1,414) of the eliminated soundings had average surface winds greater than 20 kt. The plotted 95% confidence band encloses the most probable best-fit lines given the spread of data. An increase (decrease) in PI occurs over period 0-9 LST (11-21 LST). The former interval exhibits a slightly greater best-fit slope magnitude than the latter, but contains fewer observations as well. Despite the data gap over 9-11 LST, the average PI just before 9 LST closely resembles that just after 11 LST. This suggests that PI peaks somewhere within the gap. Furthermore, uncertainty in the best-fit slope over 11-21 LST still encompasses negative values, suggesting that PI is not constant and consequently must increase over 0-9 LST. Decreasing ventilation index over 0-9 LST emanates from increasing
Figure 5-2: PI wind (a) and the ventilation index (b) are plotted diurnally for warm-water dropsondes. Best-fit lines are graphed in yellow over intervals 0-9 LST and 11-21 LST. For the PI wind, the left-hand (right-hand) best-fit slope is $1.12 \pm 1.16$ ($-0.94 \pm 0.67$) kt h$^{-1}$. For the quantity $\log_{10} \Lambda$, the left-hand (right-hand) best-fit slope is $-0.0236 \pm 0.0256$ ($0.0028 \pm 0.0147$) h$^{-1}$. All uncertainties are at 95% confidence.
The statistics include two-sided $t$-tests (Wilcoxon rank sum tests) with $p$ values for PI (ventilation index). A significant shift, at 95% confidence, occurs between distributions when $p$ values fall below 0.05.

PI and slightly decreasing $\chi_m$. The decrease in $\chi_m$ (not shown) stems from a decrease in $s^*_m - s_m$. Hence increasing PI and mid-level entropy (toward saturation) contribute to lower ventilation indices. The identified trends are better visualized in the diurnally-partitioned histograms of Figure 5-3. Diurnal shifts in mean values are assessed numerically in Table 5.1. Significant shifts occur for PI and morning-based ventilation indices.

Figure 5-4 reveals that the quantity $CAPE^*_{RMW} - CAPE_{RMW}$ significantly affects the resulting PI. The best-fit slope magnitude for $CAPE^*_{RMW}$ is greater than that of $CAPE_{RMW}$ within the two main LST intervals. The difference also maximizes somewhere in the middle gap, where PI peaks. The diurnal variation in outflow temperature $T_0$ (not shown), a variable in $CAPE^*_{RMW}$, has a slight increase in late-night hours but remains constant in afternoon/evening hours. Note that the outflow temperature of any lifted air parcel is its temperature at the LNB. The means of these two intervals however do differ significantly, with the former (latter) being $-71.1^\circ C$ ($-72.3^\circ C$). But trends in $T_0$ appear insignificant because as it increases late at night, PI decreases. Instead, we see the opposite trend in PI for late-night hours, largely from the increase in $CAPE^*_{RMW} - CAPE_{RMW}$.

More details in the PI results can be deduced from the composite sounding in Figure 5-5. Only composite data from morning are displayed since that period had a noticeable trend for PI and ventilation in Figure 5-2. The dewpoint temperature was calculated on every sounding using the relative humidity and the equation of

<table>
<thead>
<tr>
<th>LST Intervals (h)</th>
<th>$p_{Vmax}$</th>
<th>$p_{\Lambda}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-3, 6-9</td>
<td>0.051</td>
<td>0.046</td>
</tr>
<tr>
<td>0-4.5, 4.5-9</td>
<td>0.039</td>
<td>0.031</td>
</tr>
<tr>
<td>11-14, 18-21</td>
<td>0.007</td>
<td>0.819</td>
</tr>
<tr>
<td>11-16, 16-21</td>
<td>0.001</td>
<td>0.243</td>
</tr>
</tbody>
</table>

Table 5.1: The statistics include two-sided $t$-tests (Wilcoxon rank sum tests) with $p$ values for PI (ventilation index). A significant shift, at 95% confidence, occurs between distributions when $p$ values fall below 0.05.
Figure 5-3: The histograms depict diurnal shifts in normal distributions containing all PI and ventilation index data points. Some of these shifts, as the histogram binning tries to reflect, are significant and confirmed statistically in Table 5.1. The bin spacing along the horizontal axis is approximately 0.1929 in (c) and 0.1882 in (d).
Figure 5-4: $CAPE_{RMW}$ (blue) and $CAPE^*_{RMW}$ (black) quantities are for the data in Figure 5-2(a). The best-fit slope for $CAPE_{RMW}$ in interval 0-9 LST (11-21 LST) is $57.6 \pm 96.8$ (9.0 $\pm$ 53.9) J kg$^{-1}$ h$^{-1}$. The best-fit slope for $CAPE^*_{RMW}$ in interval 0-9 LST (11-21 LST) is $158.5 \pm 118.8$ ($-74.4 \pm 64.7$) J kg$^{-1}$ h$^{-1}$. All uncertainties are at 95% confidence.

Buck (1996). The temperature profile is nearly constant throughout the day as seen by the overlapping solid and dashed red curves. The same is not true, however, for dewpoint, which shows a significant increase in the troposphere from early to late morning. This trend means that the column moistens during the period. The boundary layer also experiences that trend slightly, affecting the mixing ratio of lifted parcels in PI calculations. These effects are apparent with the two lifted-parcel curves. Notice how both curves shift to a higher overall temperature from 0-3 LST to 6-9 LST. This behavior translates to a greater area enclosed between the unsaturated RMW, saturated RMW, and environmental parcel temperatures. This increase in area is proportional to the CAPE difference in Equation 5.1, leading to higher PI.

\[ T_d = \frac{257.14 \ln(H) + \xi}{18.678 - \ln(H)} - \xi, \text{ where } \xi = \frac{T}{257.14 + T} (18.678 - \frac{T}{257.14}) \text{ and } T \text{ is in degrees Celsius. It is more accurate at upper levels, where the temperature drops below the optimum range of the Bolton (1980) formula.} \]
Figure 5-5: Two sets of composite soundings are graphed for observed temperature (red), observed dewpoint (dark green), the lifted pseudoadiabat acquiring $CAPE_{RMW}$ (yellow), and the lifted pseudoadiabat acquiring $CAPE^*_{RMW}$ (magenta). The composites are averaged over interval 0-3 LST (solid), and 6-9 LST (dashed). Note that the observed dropsonde temperatures lie at the same pressure levels as the dewpoint values. Above these levels the temperature profile emanates from the composite in Figure 5-1, applied using methods described in Section 5.1. The composite pseudoadiabats below the 900-hPa level were omitted, given high variability in the lifted condensation level within the boundary layer. The underlying tephigram is identical to that of Figure 5-1.
Chapter 6

Analysis and Discussion

We now review this project’s results and compare them with prior statistical and theoretical studies. Throughout the previous three chapters, we have investigated and documented diurnal profiles of short-term wind tendencies under different physical conditions. The most significant findings included the warm-water RI events and intensity indices during a few periods of the day. We generated those results from two independent data sets. Relationships between the two sets of results may provide evidence either refuting or supporting proposed physical mechanisms that modulate TC intensity.

6.1 Observed Best-Track Rates

Results from the best-track data of Chapters 3 and 4 suggest that diurnal variability of intensification rates becomes more visible when analyzed under specific data constraints. Diurnal analysis of rates from both land and water areas produces extensive noise that hides possible trends. The procedures of this study were successful at applying constraints that limit the analysis to two broad, but tremendously different, physical situations. The noise issue, however, was not eliminated for results originating over land areas. The noise remained prominent even after using the IWDM to define a basic state for TC decay rates and generate diurnal departures. Furthermore, the daily insolation cycle has no apparent effect on land-decay rates, including
wind-tendency intervals that span sunrise or sunset. Similarly, the warm-water results exhibited large scatter in 6-h rates and did not exhibit significant diurnal shifts in the mean symmetric-wind tendency. The warm-water rates, however, are more advantageous to our study since they are abundant and follow well-defined, normal distributions. This feature allowed us to bin the data diurnally, and uncover a detectable diurnal trend in RI. The RI frequency maximizes (minimizes) during mid-morning (late-evening) hours. The warm-water data constraints ultimately isolated this trend, which otherwise appears obscured when combining all over-land and over-water 6-h tendencies for analysis.

The warm-water RI trend is the most striking result from the best-track data. The trend nearly follows the phase in the diurnal cycle of oceanic deep convection as described in Gray and Jacobson, Jr. (1977); Liu and Moncrieff (1998); Yang and Slingo (2001). The cycle in RI, although discrete at 6-h, occurs as departures from a full distribution and therefore provides some important statistical evidence. This evidence gives a broader scope to our analysis than evidence from case studies, such as those discussed in Section 2.2. Relevant case studies usually focus on specific RI events, rather than multiple events over extended periods. From our study the RI trend became visible under the data restrictions of Section 4.1, without needing others such as low wind shear or moist mid levels. More interestingly, the trend even appeared when binning all over-water data; but the result had less certainty and amplitude when graphed on the axes from Figure 4-5(b). In addition, the minimum (maximum) in RI frequency occurs over 18-24 LST (6-12 LST), when all symmetric-wind tendencies have a mean insolation of 0.40 (20.64) MJ m$^{-2}$ as seen in Figures 4-3 and A-3. This relationship, however, does not follow smoothly because for interval 0-6 LST (12-18 LST), the mean insolation is 1.61 (13.18) MJ m$^{-2}$. Therefore, we see that RI departures become positive as insolation rapidly increases, especially for intervals spanning mid or late morning.

Our results support proposed physical mechanisms that contain a mid to late morning maximum in convective intensity. Two mechanisms having this property involve either the radiatively-driven divergence in cloud-free areas (Gray and Jacob-
son, Jr., 1977), or upper-level destabilization near cloud tops (Hobgood, 1986). At this moment, however, we cannot determine which of the two mechanisms fits best with results from Chapter 4. Solving this issue would require upper-air or vertical reanalysis data, which are not included in the best-track data set. Future application of the necessary vertical data to the best-track data can refine the validity of the two mechanisms.

One drawback of the best-track results, for the sake of future diurnal analyses, is the data’s temporal resolution. The 6-h tendencies are discrete and estimating intensity readings within the 6-h intervals would require other comprehensive data collection and assimilation methods. In addition, the 6-h tendencies obviously rely on interpolation of each tendency’s two observed wind speeds. The interpolation usually involves Dvorak methods (e.g., see Dvorak 1984) to determine TC intensity from infrared data over the open oceans. We assume, despite the Dvorak-based interpolation, that no temporal error accompanies each calculated wind tendency. Most of the uncertainty in TC intensity originates from satellite-based methods and not from the relatively limited in situ observations of dropsondes (Torn and Snyder, 2012). Although dropsonde data lacks spatial coverage in individual TCs, they have a large spread in daily observation times.

6.2 Dropsonde Intensity Indices

We can analyze diurnal signals in the vertical atmospheric profiles from Chapter 5 alongside those of best-track data. The main difference here, as opposed to best-track data, is that the dropsonde results are based on theoretical quantities. Nevertheless, these quantities were found using observed soundings that give a picture of ambient conditions surrounding TCs. An advantage here is that the data is recorded in real time and, for example, not interpolated from Dvorak methods or grid techniques associated with reanalyses. All soundings included in our analysis were successfully selected according to two primary constraints: SST ≥ 27°C and low surface winds.

The data selection produced acceptable values for PI and relevant parameters.
Values of $CAPE_{env}$, though not used to find PI, reflected locations at large TC radii and away from the eyewall. Molinari et al. (2012) find from Gulfstream-IV dropsondes that pseudoadiabatic $CAPE_{env}$ is about 1500 J kg$^{-1}$ or greater at TC radii further than 400 km. They also obtained a minimum in $CAPE_{env}$ near the center of TCs. From our study, the average $CAPE_{env}$ for interval 0-9 LST (11-21 LST) was about 1930 (1680) J kg$^{-1}$. This verifies that most soundings displayed in the results originate from each TC’s outer regions. These regions are more representative of the environment’s PI, since PI tends to decrease towards a TC’s center (K. Emanuel, personal communication). In addition, as noted in Emanuel et al. (2004), observed PI varies slowly in time and highly depends on changes in SST. The composite sounding of Figure 5-5 has a surface temperature that is nearly the same between the 0-3 LST and 6-9 LST periods. Furthermore, the SST is (and should be) randomly distributed (up to a maximum of $\sim$ 30°C), and the best-fit line for dropsonde surface temperature over 0-9 LST (11-21 LST) has a negligible slope of $\sim$ 0.017 ($-0.018)^{\circ}$C h$^{-1}$ (not shown). Although probably important, we do not consider if temperature errors from incoming sunlight and instrumental defects impact the aforementioned “negligible” best-fit slope. Therefore, assuming proper instrumentation, changes in surface parameters are unlikely influencing the upward trend in PI.

The trend in PI is connected with that of $CAPE_{RMW}^*$. Note in Figure 5-4 how the fitted lines for $CAPE_{RMW}$ have similar slope and therefore an insignificant difference between intervals. This similarity however is not apparent for $CAPE_{RMW}^*$, which has a steeper slope in period 0-9 LST. The trend in $CAPE_{RMW}^*$ is not affected by how the flight missions are conducted. We argue this because over both intervals of data, the top pressure on each sounding begins near 200 hPa (at $\sim$ 0 LST and 12 LST) and ends near 170 hPa (at $\sim$ 9 LST and 21 LST). Since this trend in flight level is the same on both intervals, it is not affecting the trend in $CAPE_{RMW}^*$ for the first interval. Further confirmation from Molinari et al. (2012) shows that the average dropsonde-release pressure from the Gulfstream-IV is near 180 hPa. Hence, consistent dropsonde-release heights and appended portion from PREDICT (2010) are not modulating any CAPE values.
The results from Chapter 5 provide additional evidence supporting warm-water RI trends. For RI events, Kaplan and DeMaria (2003) determined statistically that the averaged 700 hPa to 850 hPa relative humidity is higher by 4.3% than otherwise. They also find that the PI is higher during RI events and argue that TCs are likely to undergo RI under the accompanying environmental conditions. Results from the dropsonde data support both statistical observations. Note in the composite sounding of Figure 5-5 that the dewpoint is higher later at night, in the interval 6-9 LST. The higher dewpoint implies a moister troposphere since the temperature remains relatively unchanged between 0-3 LST and 6-9 LST, and because the dewpoint increases roughly as function \( \ln(H) \). Over the period 0-9 LST, PI increases as well and seems to maximize between 9-12 LST given the increasing (decreasing) trend in the morning (afternoon) hours. The occurrences of increased moisture and PI nearly coincide with the RI maximum depicted in Figure 4-5. Although the results agree statistically with those of Kaplan and DeMaria (2003), the physical influences of high-moisture and/or high-PI environments on RI remain unknown.

Trends in the morning ventilation index are likely tied to those of PI. The 0-4.5 LST (4.5-9 LST) ventilation distribution has a mean of 0.1288 (0.1020). These means differ with 95% confidence, and are slightly higher than the climatological mean of Tang and Emanuel (2012).\(^1\) A smaller \( \Lambda \) evolves toward late morning, when RI frequency increases. Low ventilation indices represent less intrusion of dry air into TCs and better chances for maintenance and ultimately RI. PI is likely the only factor in decreasing the ventilation index, despite a slight decrease in \( \chi_m \) (from quantity \( s^*_m - s_m \)) and \( u_{shear} \) (not shown). The combined trend from those quantities produce a significant morning decrease in \( \Lambda \).

Dropsonde results for PI, ventilation, and the composite sounding of Figure 5-5 narrow the scope of underlying physical mechanisms. The most interesting PI and ventilation results occur when RI frequency increases. Over that same interval, the temperature profile remains nearly constant through the day, including at upper

\(^1\)This small disagreement is likely due to the negative bias in PI wind for soundings taken within TCs.
levels. This can be seen in Figure 5-5 for morning hours. The composite soundings do not exhibit variation in upper-level lapse rates, which seems to contradict the diurnal destabilization theory presented in Hobgood (1986). We argue this, however, only considering the large-scale conditions near or just surrounding the outer edges of TCs. We do not incorporate a TC’s entire structure since the dropsondes were selected from areas outside the eyewall. The theory of Hobgood (1986) could hold at smaller scales; an analysis of detailed soundings within convective cells may support that theory. As a result, the large-scale convergence mechanism of Gray and Jacobson, Jr. (1977) may play a role in producing the apparent results. In addition, theory from Tao et al. (1996) applies to our morning trend in atmospheric moisture (see Figure 5-5). Since the atmospheric column moistens during late-night hours, large-scale convergence under the TC can both increase lift and enhance latent heating within the TC core. Greater low-level convergence and moister mid-levels during late-night hours constitute an important dynamical trend that seems to support the results.
Chapter 7

Conclusion

We have uncovered observational evidence that sheds some light on diurnal variability in short-term TC intensity trends. Despite the randomness of general over-land and over-water wind tendencies, the most noticeable diurnal oscillation appeared for warm-water RI events. We emphasize that the RI trend accounts for increasing occurrences of the RI rates, rather than increasing mean RI rate, during morning periods. Our statistical result supports a prediction mentioned in Tripoli (2006), which states that diurnal signals become more apparent during periods of intensification and enhanced vertical convective activity. Such periods occur in morning hours, and the physical mechanisms we have discussed give large-scale explanations for the apparent maximum in convection and RI frequency. The mechanisms we accept here, described in Gray and Jacobson, Jr. (1977) and Tao et al. (1996), generally describe the diurnal TC physics that contribute to the apparent trends. A more detailed analysis of small-scale convective features, however, could support or refute those mechanisms. Finally, the RI trend has been tested against trends in PI and ventilation index, and fulfills statistical relationships recorded in Kaplan and DeMaria (2003) for Atlantic TCs. Diurnal analysis of the intensity indices gave more temporal and vertical details supporting our accepted physical mechanisms.

Many physical interactions that modulate intensity diurnally remain a mystery, and the subject requires additional research. Although our findings include a trend for RI, they lack information for physical situations that exhibited no diurnal trends. In
addition, not knowing all physical factors that produce RI is arguably the most problematic aspect of short-term intensity forecasting. Hence, given that RI is anomalous, diurnal model simulations of RI in TCs would yield more details that one can use to craft better methods in future observational analyses. Most importantly, as we have argued, those simulations would depend highly on surrounding physical influences that should be defined carefully. Diurnal influences on rapid decay, be it over land or water, also requires consideration in future studies on TC physics. On the other hand, our dropsonde data was not quality-controlled according to NOAA HRD. Therefore an analysis of trustworthy vertical data is necessary to verify our resulting diurnal trends in PI and ventilation index. Finally, one could rigorously test those trends’ magnitudes in model simulations and investigate their physical implications for RI and rapid decay.
Bibliography


Appendix A

Insolation Profiles

This Appendix documents results of top-of-atmosphere insolation calculations for intensification rates. This information is given for land-decay MSSW tendencies in Figure A-1, for warm-water MSSW tendencies in Figure A-2, and for warm-water symmetric-wind tendencies in Figure A-3. These calculations only apply for rates since they occur over extended periods (e.g. 6 h) and therefore can be applied to Equation 3.3. In addition, the information presented here is supplementary in nature because it does not capture overnight trends. The insolation is usually zero for nighttime rates, and the data arrangement in the following figures constricts spread in the nighttime rates. In each of the figures, subfigure (b) arranges data such that it distinguishes insolation between the first and second halves of the day. Note how the insolation reaches a maximum for warm-water tendencies centered at solar noon. For land-decay (warm-water) rates, the insolation on the horizontal axes of (b) subfigures begin at approximately 6 LST (8 LST) and proceeds forward through the day. This arrangement in insolation however yields no significant diurnal trend. This conclusion applies to running averages of the wind tendencies (not shown), which remain slightly above zero near $\sim 0.3$ kt h$^{-1}$ with no significant deviations.
Figure A-1: The insolation for land-sector MSSW tendencies are arranged diurnally in (a), similar to the format of Figure 3-6. The land-decay rates in (b) are plotted as a function of the diurnal variability in insolation. The left (right) vertical dashed line is centered at the maximum (zero) insolation which occurs near solar noon (at night). From left to right, the three LST intervals bounded by the dashed lines are approximately as follows: $6 < \mathcal{T} < 12$, $12 < \mathcal{T} < 24$, $0 < \mathcal{T} < 6$. Most nighttime points contain zero insolation.
Figure A-2: Same as Figure A-1 but for warm-water MSSW tendencies. The left dashed line is centered at the maximum insolation of 26.174 MJ m$^{-2}$. From left to right, the three LST intervals bounded by the dashed lines are approximately as follows: $8 < \mathcal{T} < 12$, $12 < \mathcal{T} < 24$, $0 < \mathcal{T} < 8$. 
Figure A-3: Same as Figure A-2 but for symmetric-wind tendencies.