# The Role of Convective Available Potential Energy (CAPE)

# in

# Tropical Cyclone Intensification

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Abstract

The formation and intensification of tropical cyclones is still a work in progress. For more than 50 years several theories have been put forward to try and explain how these atmospheric phenomena grow and maintain itself from dissipation. Two competing theories -CISK and WISHE - have emerged where each has its strengths and weaknesses. CISK which stands for conditional instability of the second kind, states that the vortex develops in a conditionally unstable environment where a positive feedback loop develops between the frictional convergence in the boundary layer and the release of latent heat in the convective updraughts. WISHE which stands for wind induced surface heat exchange explains that the growth and development of the cyclone is driven exclusively by the surface fluxes of heat and moisture from the underlying ocean and there is no contribution by ambient conditional instability. The atmosphere is moist neutral to both upright and slantwise convection. In more recent times some within the TC community have neither supported the WISHE nor CISK paradigm as the answer to intensification and they have put forward a new paradigm which is based on the existence of towering clouds called vortical hot plumes, that form in the inner core of TCs and add to its system scale circulation. Although many researchers have studied the intensification from many different angles and perspectives none have been able to adequately answer the question of how does CAPE (convective available potential energy) fit into the theory of intensification. Although, Rotunno and Emanuel mentioned that in WISHE there is no initial CAPE, due to the fact that CAPE can be generated by evaporation at the sea surface, it is still not clear what role CAPE plays during intensification. To better understand the role CAPE plays during the intensification process we use a hierarchy of models with different complexity. We first develop a conceptual model which explains the role of CAPE in intensification. CAPE is generated by surface fluxes in the boundary layer. Its presences causes high entropy air to be carried to the eyewall where latent heat is released and enhances the negative radial temperate gradient which in turn enhances the inflowing air through the increase of the gradient wind. We use a simple low-order box model to further our understanding. When we damp CAPE in our experiments this hinders the intensification because the secondary circulation in the absence of CAPE transports low entropy air into the eyewall. Other researchers have proven that the secondary circulation plays a vital role in the balancing of energy production in the boundary layer. The next group of experiments were carried out in using a modified version of the Ooyama three-layer model where unbalanced dynamics are employed. Similar results were obtained with the Ooyama model. We were able to successfully prove that the WISHE mechanism for intensification cannot fully account for intensification. The experiment that obtained the highest amount of intensification was the one that triggered convection through frictional convergence. Even when we included the efficiency

precipitation parameter, a parameter that is supposed to put further restrictions on the surface flux induced heating so that intensification can occur under the WISHE paradigm, failed to produce a cyclone. Our last group of experiments study CAPE by a series of sensitivity tests where a non-hydrostatic cloud resolving model is used. We run experiments for both the axisymmetric and 3D configurations where the initial CAPE values differ, similar to the previous experiments.Then we test to see how varying amounts of CAPE generated by the different values for the surface transfer We see that the results in the axisymmetric are generally more intense and yield more robust results. The 3D wind speeds tend to be weaker. In the case for surface transfer enthalpy, all the experiments produced tropical cyclones even when the parameter was reduced to quarter its value. From our experiments we support the fact that a radial gradient of CAPE is needed to support intensification and as a result it is an important parameter in intensification.

Zusammenfassung

Die Entstehung und Intensivierung von tropischen Zyklonen sind immer noch Gegenstand der Forschung. Vor mehr als 50 Jahren wurden mehrere Theorien vorangetrieben, um zu verstehen wie diese atmosphärischen Phänomene sich verstärken und erhalten können trotz der vorhandenen Dissipation. Zwei konkurrierende Theorien – CISK und WISHE – kamen auf, wobei jede ihre Stärken und Schwächen aufweisen. CISK, was bedingte Instabilität zweiter Art ("conditional instability of the second kind") bedeutet, besagt, dass der Wirbel sich in einer bedingt feuchtinstabilen Umgebung entwickelt, wobei eine positive Rückkopplung zwischen reibungsbedingter Strömungskonzergenz in der Grenzschicht und Freisetzung von latenter Wärme in den konvektiven Aufwinden wirkt. WISHE, was für windinduzierter Oberflächenwärmeaustausch (wind induce surface heat exchange) steht, erklärt, dass die Entstehung und Intensivierung der Zyklone ausschließlich durch Flüsse von sensibler und latenter Wärme von dem darunterliegenden Ozean verursacht wird, so dass bedingte Feuchtinstabilität der Umgebung keinen Beitrag liefert. So ist bei dieser Betrachtung die Atmosphäre feuchtneutral in Bezug auf senkrechter und schräger Konvektion. In jüngster Zeit wurde von einigen Wissenschaftlern weder das WISHE noch das CISK Paradigma als Intensitätsmechanismus anerkannt. Sie entwickelten ein neues Paradigma, welches auf die Existenz von tiefen Konvektionswolken, die als warme Wirbelplume ("vortical hot plume") bezeichnet werden, sich im inneren Kern von tropischen Zyklonen bilden und zur Zirkulation des Gesamtsystems beitragen. Obwohl viele Wissenschaftler die Intensivierung aus verschiedenen Blickwinkeln und Perspektiven studiert haben, hat noch keiner die Frage adäquat beantwortet, wie die Existenz von konvektiv verfügbarer potentieller Energie (CAPE, "convective available potential energy") in die Theorie der Intensivierung passt. Obwohl Rotunno und Emanuel erwähnten, dass in WISHE keine anfängliche CAPE vorhanden ist, ist es immer noch unklar welche Rolle CAPE bei der Intensivierung spielt, da CAPE durch Verdunstung an der Wasseroberfläche generiert werden kann.

Um die Rolle von CAPE bei der Intensivierung besser zu verstehen, nutzen wir eine Hierarchie von Modellen verschiedener Komplexität. Wir entwickeln zunächst ein Konzeptmodell, dass die Rolle von CAPE erklären kann. CAPE wird durch Oberflächenflüsse in der Grenzschicht generiert. Seine Existenz verursacht, dass Luft mit hoher Entropie zum Augenwall transportiert werden kann, wo latente Wärme freigesetzt und den negativen radialen Temperaturgradienten verstärkt, welche wiederum das Einströmen aufgrund des höheren Gradientwindes verstärkt. Wir benutzen ein einfaches niedrigdimensionales Boxmodell, um das Verständnis weiter zu vertiefen. Wenn wir CAPE dämpfen in unseren Experimenten, dann behindert das die Intensivierung, weil die Sekundärzirkulation in der Abwesenheit von CAPE Luft mit niedriger Entropie in den Augenwall

transportiert. Andere Wissenschaftler haben bewiesen, dass die Sekundärzirkulation eine entscheidende Rolle für die Energiebilanz in der Grenzschicht spielt. Die nächste Gruppe der Experimente wurden mit einer modifizierten Version des Dreischichtenmodells von Ooyama durchgeführt, wobei unbalanzierte Dynamik verwendet wird. Ähnliche Resultate wurden mit dem Ooyama-Modell erzielt. Uns gelang es zu beweisen, dass der WISHE-Mechanismus nicht vollständig die Intensivierung erklären kann. Das Experiment mit dem höchsten Ausmaß an Intensivierung war das, in welchem Konvektion nur durch Konvergenz in der Grenzschicht ausgelöst wird. Selbst, wenn wir einen Effizienzparameter – ein Parameter, der Einschränkungen auf die durch Oberflächenflüsse induzierte Erwärmung beschreibt und für die Erklärung von WISHE wichtig ist – einführen, stellen wir immer noch keine Zyklonenentstehung im Modell fest. Unsere letzte Gruppe von Experimenten untersucht den CAPE-Einfluss durch eine Reihe von Sensitivitätsexperimenten mit einem nichthydrostatischen wolkenauflösenden Modell, wobei wir die Experimente sowohl in axialsymmetrischer als auch in 3D Konfiguration durchführen. Die anfängliche CAPE unterscheidet sich wie bei den vorangegangenen Experimenten und dann testen wir, wie verschiedene Werte des Oberflächentransferkoeffizienten für Enthalpie die Intensivierung beeinflussen. Die Windgeschwindigkeiten bei den 3D-Experimenten weisen geringere Werte auf als im axialsymmetrischen Fall. In den Fällen für den Einfluss des Oberflächentransfers von Enthalpie entsteht immer eine tropische Zyklone, sogar wenn der Transferkoeffizient auf ein Viertel des Referenzwertes reduziert wird. Die Experimente stützen die Hypothese, dass ein radialer Gradient von CAPE benötigt wird, um Intensivierung zu erklären und als Ergebnis erhärtet sich, dass CAPE eine wichtige Größe bei der Intensivierung darstellt.

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# Chapter 1

# Introduction

Tropical cyclones rate as one of nature's most deadly and most costly disasters. These highly destructive storms can cripple economies and bring the death toll into the hundreds of thousands (Anthes 1982). They rate as one of nature's most destructive natural disaster where the major causes of all these damages are from the strong winds, storm surges and heavy precipitation (Anthes 1982). Interestingly, they also play a very critical role in the earth's energy budget, a fact many tend to forget. They help to maintain the earth's energy balance. Additionally, the copious amount of rain that tropical cyclones bring help to maintain the hydrological balance of a region (Anthes 1982). As a result, an accurate forecast of these storms is pertinent so those who are most vulnerable can be better prepared for the storms. Tracking the path of the cyclone and its intensity are still very active areas of research. Due to the challenges that arise when trying to forecast intensity change much interest and attention is being devoted to a better understanding of intensity change (Montgomery and Smith 2017).

## 1. Classification, Climatology and Formation of Tropical Cyclones

A tropical cyclone is the generic name given to any low pressure system that develops over warm tropical ocean waters. Depending on their maximum wind speed ($V_{max}$) and location they are named accordingly. If $V_{max} \leqslant 17$m/s, it is called a tropical depression. If $17$m/s $< V_{max} < 33$m/s, it is called a tropical storm and if $V_{max} \geqslant 33$m/s, it is called a hurricane if it is formed in the western North Atlantic and eastern Pacific, typhoons if it formed in the western North Pacific (Emanuel 2003). Annually, approximately 80 tropical cyclones form over the warm tropical oceans but only about 60% ever make it to maturity (Emanuel 2003). For a these systems to develop certain conditions need to be present. Gray (1968, 1979) developed a set of climatological conditions that need to be present for the formation of a cyclone but are not sufficient - in other words even if these conditions were present a TC still may not form due to other factors. These conditions are:

i. Sea surface temperature must exceed 26 °Celsius with an ocean depth of at least 60m
ii. There needs to be a negative vertical gradient of equivalent potential temperature in the atmosphere
iii. The relative humidity of the middle troposphere must be at least 40%
iv. Large values of the Coriolis parameter
v. There must be large values of low-level absolute vorticity
vi. The vertical shear of the horizontal wind must be mild

The first three parameters are intimately related to each other because they set the environmental conditions that will support deep convection. As Palmén (1948) in his research noted that tropical cyclones can only develop in tropical waters at a critical temperature, and below it, formation would not be possible. As a result of this critical value, there will be only certain regions where the conditions are favorable for TC formation. Anthes (1982) noted that no TCs form in the southwest Atlantic and the northeast Atlantic because of the cooler temperatures. The vertical gradient of equivalent potential temperature needs to be negative and large so that there will be strong convective instability, and due to the entrainment of middle level air, the moisture content in this region needs to be high enough to support the release of latent heat from the converging air. The next two parameters are also related to each other. They set the the conditions to ensure the vortex maintains rotation and life. It has been observed that TCs do not form at or near the equator. They do not even cross the equator. This is because in order to have rotation, a necessary condition for a vortex, there needs to be a non-zero large enough Coriolis parameter so that the converging air will have a rotation flow supporting a strong vorticity (Anthes 1982). Anthes (1982) pointed out that any convective system that forms on the equator tends to be divergent and not rotational. The last condition has more to do with the way the cyclone interacts with its environment. TCs are actually large-scale eddies that are embedded in an environment that has its own flow and circulation. If the wind shear is too great all the moisture and temperature throughout the cyclone can be advected away causing the cyclone to dissipate (Anthes 1982). The cyclone needs a sufficient supply of moisture and heat to survive.

## 2. Structure of Tropical cyclones

Through satellite images, reconnaissance aircraft data, groundbased and airborne data we are able to see the structure of a tropical cyclone (Emanuel 2003). Three-dimensional pictures of TC have been developed by compounding of data. Through these pictures we are able to see the detail structure of a TC. Figure 1.1 shows a sketch of a mature tropical cyclone with its flow features. In the centre as shown in the figure there is a cloud-free eye where the air is subsiding as the red arrows indicate. Surrounding the eye are thick tall towering clouds that have an amphitheater shape to them. This region is the eyewall. As indicated by the thin blue arrows, the air rises here in a thermally direct circulation. It is moist here and the convection is most intense. This region is where the radius of maximum winds are located. The most intense precipitation and wind speeds are here, also. Beyond this region are smaller clouds that have less intense precipitation and they form the trailing rainbands. There are two circulations present in a TC: the primary circulation and the secondary circulation. The primary circulation is confined to the horizontal plane and it flows cyclonically (northern hemisphere). The flow in the primary circulation in a TC is known to be in gradient wind balance except in the boundary layer where it is in supergradient due to the imbalance of forces (Willoughby 1988).The balance of the three forces are: the pressure gradient force, the centrifugal force and the Coriolis force

$$\frac{1}{\rho}\frac{\partial p}{\partial r} = \frac{v^2}{r} + fv \tag{1.1}$$

where $\rho$ is density, $p$ is the pressure, $r$ is the radial distance from the centre, $v$ is the tangential velocity component and $f$ is the Coriolis parameter. As one moves away from the centre of the storm the winds increase rapidly until it reaches a maximum and drops off quickly as one moves away from the maximum. That maximum is called the radius of maximum wind. According to Emanuel (2003) the winds decay following a $r^{-\frac{1}{2}}$ decay law when near the radius of maximum but faster at larger radii. The secondary circulation is the transverse circulation or the air that flows inward toward the storm centre then up into the eyewall and then out near the tropopause. This is also called the in-up-out circulation. The Sawyer-Eliassen equation describes the flow in the transverse circulation. The large-scale flow in a TC is constrained by it being in thermal wind balance, a fact that is common in large-scale flows (Montgomery and Smith 2014). The TC maintains its rotation due to the approximate conservation of absolute angular momentum above the boundary layer.

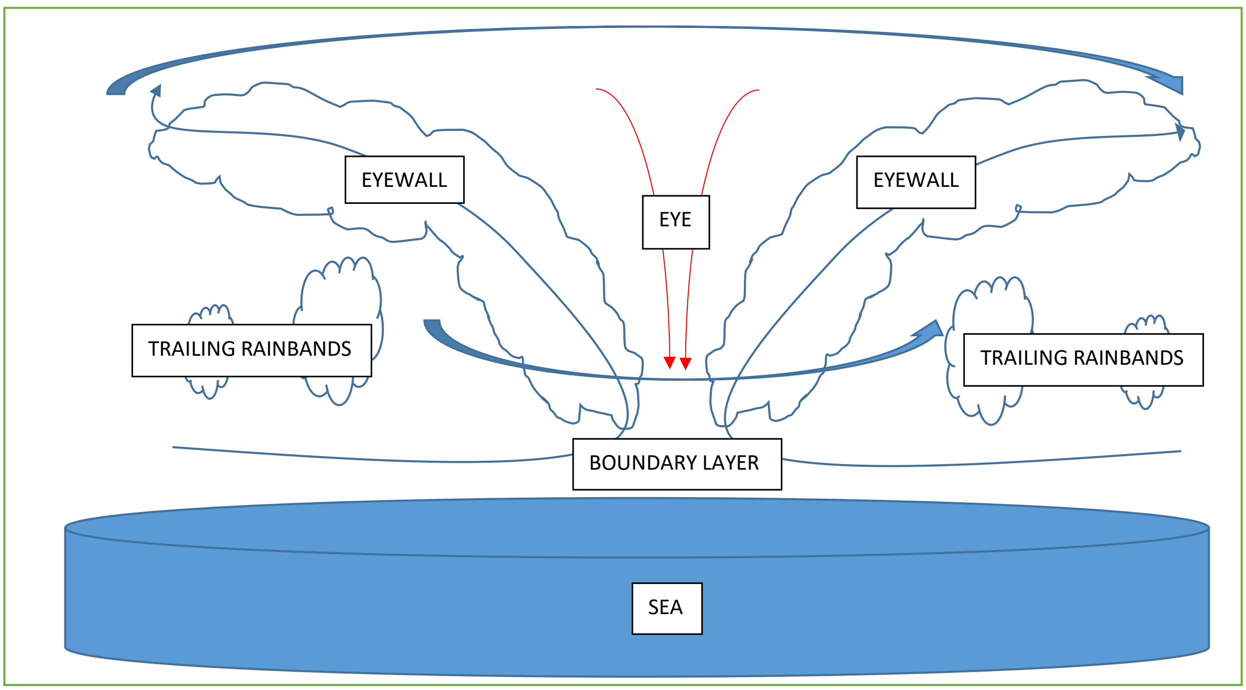



Figure 1.1: Sketch of the structure of a mature tropical cyclone showing the primary and secondary circulation with the airflow.

# 3. CAPE

CAPE is an acronym for convective available potential energy. This is the energy that is available for convection. The American Meteorological Society gives the following definition in their glossary of meteorology: "The maximum buoyancy of an undiluted air parcel, related to the potential updraft strength of thunderstorms."[1] In parcel terminology it gives a measure of how much kinetic energy a parcel would gain if it were raised a specific height in the atmosphere.

It is commonly used to give a measure of conditional instability of the atmosphere.Yano et al. (2005) argues it is better to interpret CAPE as a measure of convertibility of potential energy to kinetic energy. Convection is the mass movement of a fluid which carries with it heat energy to areas that are deficient. Convection helps to achieve stability due to the imbalance of heat distribution in the atmosphere. Yano et al. (2005) states that CAPE can also be viewed as a measure of the buoyancy force to convert potential energy of the air into kinetic energy of the rising air parcel. This view of CAPE is the Lagrangian approach where we follow the trajectory of CAPE as opposed to the Eulerian approach which evaluates CAPE at certain areas. CAPE is generated by evaporation. In tropical cyclones surface fluxes over the sea generate CAPE.

# 4. Motivation and Outline

The formation and intensification of TCs is a work in progress. Even though theories have existed in excess of 50 years we still do not fully understand the dynamics and thermodynamics that govern the TCs growth and development. For the longest while the research has led to two competing theories: conditional instability of the second kind (CISK) and wind induced heat exchange (WISHE). CISK was put forward by Charney and Eliassen (1964) while WISHE was put forward by Emanuel (1986) which later developed it into a mathematical model by Emanuel (1997). Many numerical models have been developed and used to study tropical cyclogenesis, and they have done a great job in realistically simulating the complete life cycle of a TC. As a result, numerical models can be used to determine which theory is the one that can explain intensification. Craig and

[1] `http://glossary.ametsoc.org/wiki/Convective_available_potential_energy`

Gray (1996) studied these two theories and came to the conclusion that enthalpy transfers increase the intensification rate but not the surface drag which is essential for cisk. This result argues for the WISHE theory. However, Montgomery et al. (2009) was able to prove in a three-dimensional model that the wind speed dependence surface fluxes were not essential, although they play a key role in the WISHE theory.

The WISHE paradigm assumes that the atmosphere is always neutral to CAPE because convection acts so fast. Gray and Craig (1998) and Frisius (2006) have investigated to see if that assumption holds true and in doing so they adopted an efficiency parameter which ensures that the radial gradient of heating is negative. There is justification for the introduction of this parameter. Frisius (2006) has shown that without this a cyclone cannot intensify by WISHE. Many have shown that CAPE is non-zero in TCs (e.g Frisius and Schönemann (2012), Frisius and Hasselbeck (2009), Molinari et al. (2012)). The premise of the WISHE theory is that initial CAPE is not necessary as stated by Rotunno and Emanuel (1987) but what is not certain is if CAPE has an important role during intensification since it can be generated at the surface by evaporation.

This study tackles the question whether CAPE helps to intensify the TC. CAPE creates a negative vertical gradient in the lower atmosphere. The air-sea interaction causes entropy to increase towards the centre of the cyclone and thus causing CAPE to be generated. When air rises in the eyewall it conserves its entropy so if there is a large negative gradient of saturation entropy it will rise in the eyewall. If there were no CAPE then the radial gradient would be weaker resulting in slow or no intensification. In this view CAPE is behaving like the efficiency parameter. Some studies have viewed tropical cyclogenesis as a combination of both CISK and WISHE (e.g Schönemann and Frisius (2012)) but there is no evidence to support this. In this thesis we focus on investigating the role CAPE plays in tropical cyclogenesis by using a hierarchy of models. The simpler models gives us the ease of understanding the mechanism while the more complex models helps to validate the mechanism. The thesis is set up where chapter 2 gives a brief summary of the more well-known intensification theories, chapter 3 presents results of a low-order model, chapter 4 uses the modified Ooyama model, Chapter 5 uses CM1 to run sensitivity tests in both axisymmetric and 3D and the final chapter 5 provides the conclusion.

# Chapter 2

# Different approaches to Intensification

## 1. Introduction

The term tropical cyclone (TC) intensification can draw much debate and contention. Presently, there is no one conclusive theory that can explain the dynamics and thermodynamics of a developing storm. The mechanisms and processes that are at play when a depression transitions to become a full blown Hurricane or Typhoon can vary depending on which theory one ascribes to. Even the term intensification can be problematic to define so for simplicity when we speak of intensification we are referring to the tangential increase in winds or as some authors may call it, the spin-up of TCs. The tangential wind is the wind field in the primary (horizontal) circulation. Montgomery and Smith (2014) highlights the fact that even the theory explaining intensification needs to be properly developed where all the major fluid dynamical and thermodynamical processes are included so that we can better understand our model forecasts and also be able to identify the important process that we may need to better understand. When researchers are studying TC intensification, they observe the minimum surface pressure drop in the eye and the maximum wind speed in the eyewall. These two parameters indicate the intensity of the storm and as a result are very significant parameters when studying TCs. Since there is a lack of collective agreement on how a TC develops, this chapter presents a brief outline of the four major schools of thought on intensification and their limitations or drawbacks. In the last section a conceptual model of intensification which acknowledges and includes CAPE/SCAPE provides the framework for our understanding of CAPE's role in intensification. The conceptual model section was taken directly (word for word) from Lee and Frisius (2018).

# 2. Intensification Theories

*a. CISK*

Conditional instability of the second kind, commonly known as CISK was a theory put forward by Charney and Eliassen (1964) to explain the evolution of a pre-hurricane depression in a conditionally unstable atmosphere. In their highly acclaimed paper, their theory viewed TC development as a positive feedback between the initial horizontal large-scale disturbance with moist convection. Instead of the pre-hurricane depression and the cumulus cells competing for the same energy, they were, in fact, working in cooperation with each other. One of the more remarkable facets of this intensification theory is the role surface friction plays. They presented this novel idea that surface friction indirectly amplifies the vortex. They further explained that friction has two functions: it dissipates the kinetic energy, and it adds latent heat energy to the system due to frictional convergence. The premise of the theory is that the cumulus cells provide the heat energy through the release of latent heat which drives the large-scale disturbance, and the disturbance provides the moisture needed for the cumulus cells by the frictional convergence in the moist boundary layer. This theory gained much success and popularity for decades where it even spawned many variations to it such as the wave-CISK instability where the moisture supply is due to internal waves and not frictional convergence (Lindzen 1974). However, over time some within the TC community had concerns with this linear instability theory and the several variations of it that came after.

The first greatest critic of CISK was Ooyama who interestingly developed a similar theory for TC intensification (see Ooyama (1964)) right about the same time as CISK. In his paper, Ooyama (1982) most noted concern was the closure assumption used for moist convection. He argued that CISK relating the vertically integrated moisture convergence to the growth of the moist convection can only be valid once the cyclone is in its mature stage. During the early stages of development due to the vast differences in the scales of motion, this cannot be possible. Both Emanuel (1994), and Raymond and Emanuel (1993) had concerns about the closure assumption. They stated that CISK also violates the principle of causality because it implies a statistical equilibrium of the water supply. In essence, the closure assumption is stating that convection consumes water as opposed to energy at the rate at which it is supplied by the macro-scale disturbance. The closure assumption was not the only source of disapproval from the community. In fact Ooyama (1982) was highly critical of the many variations of CISK that came after Charney and Eliassen. He wrote that all these other CISK parameterisations are really convection in disguise. Even the Wave-CISK by (Lindzen 1974) he noted should not even be considered a CISK theory. Another source of concern was surface fluxes of heat and moisture, an activity that is vital to the survival of a TC. Emanuel et al. (1994) had grave concerns over the omission of them. The lack of acknowledging this important process implies that TCs can develop anywhere regardless of it is land or water. In both companion papers, Emanuel (1986) and Rotunno and Emanuel (1987) have acknowledged that the source of heat and moisture that drives the storm comes from the ocean underlying the disturbance. The moisture content of air converging in the boundary layer will never be sufficient to support the convective processes if there is no ocean as its source. Montgomery and Smith (2014) stated that this was an unfair critique from Emanuel et al. (1994) especially since Charney and Eliassen (1964) did state that they are assuming that the intensification is happening over the ocean.

### *b. Cooperative Intensification*

The cooperative intensification mechanism for TC development as described by Ooyama (1969, 1982) is a cooperative process between the cumulus convection and that of the large-scale disturbance, or as Ooyama (1997) put so succinctly, it is a cooperation between the primary circulation and the secondary circulation. Ooyama (1969) was the first to successfully simulate a tropical cyclone to maturity. This was groundbreaking. Although he was successful in capturing the elements of TC intensification, he noted that there are limitations on what can be investigated. To simulate the intensification process the numerical model was designed where the atmosphere is divided into three layers: the boundary layer, the middle troposphere and the upper troposphere. The vortex, like CISK, develops in a conditionally unstable atmosphere where it maintains axisymmetry while the flow is in gradient wind balance. Air converges in the frictional boundary layer in the region of organised convection below the eyewall to maintain the warm core structure and the convective instability in the middle troposphere that helps to drive the secondary circulation. The moisture supply in the inflowing air needs to be considerably high to maintain the warm core. The organised convection induces a secondary circulation with the inflow being the first leg, the updraft in the eyewall being the second leg and the final leg is the outward flow in the tropopause. If the inflowing air in the boundary layer brings in more absolute angular momentum than what is lost to friction this will cause the swirling winds to increase causing a further drop in pressure. This drop in pressure will cause a further increase in the inflowing air which will make the convection more intense. Ooyama (1982) highlighted the fact that the strength of the primary circulation was the reason why the TC intensifies and not the frictional inflow which is driven by the radial pressure gradient. The warm air aloft that maintains the low sea-level pressure in the eye needs to held in place by a deep layer of the cyclone. The deep-layer inflow is responsible for intensification. Entrainment of air which carries angular momentum in the updraft can only be achieved if the moist updraft is convectively unstable so both air from the boundary layer and the middle layers can be lifted to the outflow level. As a result of this the vortex contracts and intensifies. To ensure that this convective instability is maintained there must be heat input into the frictionally-induced inflowing air.

The main difference between the CISK and Ooyama model is that the latter is nonlinear and includes surface heat fluxes. The other differences are minor in my view. We found in our paper (Frisius and Lee (2016)) that the linearized Ooyama model produces results similar to that of Charney and Eliassen. The crucial thing is that the nonlinear feedback (downdraughts) lead to a decrease of entropy in the boundary layer that rapidly stops growth. Only surface heat fluxes can compensate this nonlinear damping.

### *c. WISHE*

The air-sea interaction theory by (Emanuel 1986) and Rotunno and Emanuel (1987) (will be referred to as E86 and E87 thereafter, respectively) which would later be called wind induced surface heat exchange (WISHE) (Emanuel 1991), viewed the TC as an instability due to the thermodynamic disequilibrium between the underlying ocean and the atmosphere above (E86, E87). The main tenants of this theory is that the intensification of the cyclone is driven and maintained exclusively by the surface fluxes of heat and moisture from the underlying warm ocean (RE86). Unlike CISK, convection does not play a significant role. Convective processes act so fast that the atmosphere is kept in a constant state of neutrality as explained in E86. The transfer of the heat energy from the ocean in the form of evaporation is highly dependent on the wind speed. This is the primary mechanism that allows the TC to intensify under the WISHE paradigm. An increased surface wind will have an increased heat transfer which will deepen the storm which in turn will cause an even more increase in surface winds, and the cycle repeats itself (Emanuel 1991). This theory stresses the importance of surface heat transfer on TC intensification. The air that flows in the boundary layer is in gradient wind balance and when it exits this layer, it flows upward in the eyewall along surfaces of constant angular momentum conserving its equivalent potential temperature until it reaches the tropopause where it flares out to large radii and cools down due to radiative cooling. The ascent in the eyewall is neutral to both buoyancy and centrifugal forces, thus making it neutral to both upright and slantwise convection. Therefore ambient CAPE has no contribution to the intensification of the cyclone. Montgomery et al. (2009) and Montgomery and Smith (2014) succinctly explain how the tangential winds increase in E86. They state that an increase in specific humidity in the inner core region will cause there to be an increase in equivalent potential temperature in both the boundary layer and in the core warm core aloft. This results in the core warming and due to the fact that the cyclone is in thermal wind balance there will be an increase in the gradient wind at the top of the boundary layer by . This increase is transmitted throughout the boundary layer by the enhanced horizontal pressure gradient, resulting in an increase in the inflowing surface air. This increased inflow in the air will increase the surface fluxes of moisture which will lead to an increase in the specific humidity. Emanuel (1997), Gray and Craig (1998), Frisius (2006), Emanuel (2012) were all able to show that TCs can intensify without CAPE. The models that these studies used, employed many assumptions that could be misleading.

Although, E86 has been openly critical about Ooyama (1969) and its closure for convection, Montgomery and Smith (2014) have found that the two theories are more similar than one would realise. The spin-up above and in the boundary layer of Emanuel (1997) is in agreement with Ooyama's cooperative intensification theory for those same parts of the cyclone. With regards to the differences here I present them, as documented in Montgomery and Smith (2014). Firstly, the convection parameterisation schemes are different because in E86 there is pseudo-adiabatic ascent along sloping lines of constant angular momentum but in Ooyama (1969) there is upright deep convection. Secondly, E86 explicitly acknowledges the close relationship surface fluxes have with the surface wind speed, while Ooyama (1969) does not show that relationship, however he does acknowledge the surface fluxes contribution to the intensification process and its importance. Finally in the improved model of E86, Emanuel (1997) acknowledges the impact the convective downdraughts have on the thermodynamics of the boundary layer by introducing a parameter to account for these processes. Ooyama (1969) never mentions this in their intensification paradigm.

There are several drawbacks with the WISHE paradigm. The introduction of the $\beta$ parameter in Emanuel (1997) is to crudely represent the effects of convective downdraughts. This parameter cannot be derived from specific processes but only acknowledge the effects of importing low entropy air into the boundary layer. Another problematic area is the fact that WISHE assumes gradient wind balance everywhere including the boundary layer. Montgomery and Smith (2014) rejected that notion. To have an inflow of air means there must be an imbalance of forces to create an inward flow. Questions arose whether surface fluxes are the only necessary factors in intensification. Montgomery et al. (2009) was convinced otherwise. They did not think the evaporation wind-speed mechanism was an essential process. This was proven to be true in an earlier study by Sang et al. (2008) where they investigated the importance of the fluxes of latent and sensible heat, and fluxes of moisture. In that study, intensification was still able to proceed with only a minor reduction when the surface heat fluxes were capped. Frisius (2006) already detected this in his simple zero potential vorticity model. These findings shed new light on WISHE since they challenge the very core of the theory. These drawbacks have led to believe that both WISHE and CISK are not the right answer to intensification. Sang et al. (2008), Montgomery et al. (2009), Montgomery and Smith (2014) and Montgomery et al. (2015) are in support of the new rotating convective paradigm.

### *d. VHTs and The Rotating Convective Paradigm*

The vortical hot tower (VHT) route to tropical cyclone intensification otherwise known as the deep rotating convective paradigm is unique in that it does not base its theory on axisymmetric dynamics. The term vortical hot tower was first coined by Hendricks et al. (2004) to describe the cyclonically-rotating structures that form in the inner core of a TC where they contribute to the vortex circulation (Montgomery and Smith 2014). Due to their presence in three-dimensional numerical model simulations, documentation of them in observations are being made where two studies (see Reasor et al. (2005) and Sippel et al. (2006)) have already taken note of them (Montgomery and Smith (2014)). Sang et al. (2008) highlighted the fact that there is growing interest in the role asymmetries play in all stages of a TC life cycle. Montgomery and Smith (2014) bring attention to the fact that from satellite images TCs are not axisymmetric except in the inner core region of an intense storm when it's in its mature stage. Asymmetries or the asymmetric features that are present in TCs affect the intensification. These structures, which are irregularly distributed around the eye of the cyclone in the eyewall or the inner core of the cyclone add to its large scale circulation. Montgomery and Smith (2014) explains that these convective structures draw air in towards the axis of rotation while also drawing in absolute angular momentum. They further provide a description for an azimuthally averaged view of the new paradigm. In it there are two spin-up mechanisms: one in the boundary layer and the other above the boundary layer. In the case of the spin-up above the boundary layer, the air converges importing absolute angular momentum where it is conserved. This flow is driven by the inner-core convection. In the case of the air that flows in the boundary layer, the converging air also transports absolute angular momentum but it is not being conserved due to friction, and the air is in gradient wind imbalance. The converging air in the boundary layer travels a much further radial distance than the one above the boundary layer. The two spin-up mechanisms are coupled to each other by boundary-layer dynamics.

Montgomery et al. (2009) showed that CAPE was present during the intensification process but we are still unsure about the final spin-up phase where the cyclone attains maturity and forms an eyewall. Clearly, CAPE is an important element in the initial phase of intensification where the VHTs spin-up and then eventually merge. However, we are unsure of how those deep rotating convective structures can support two spin-up mechanisms in the axisymmetric view of the theory as presented by Montgomery and Smith (2014). If the VHTs are located outside the RMGW (radius of maximum gradient wind) then there will be a spin-down of the decrease in the gradient wind. This has already been proven by Frisius and Hasselbeck (2009) in a cloud resolving model where they noted that convective processes outside the eyewall delays the development of the cyclone. It is critical where these structures are located in the developing storm.

# 3. Conceptual Model of Intensification

This section was taken from Lee and Frisius (2018).

"We describe a conceptual model which highlights the key processes of intensification. The model corresponds to the azimuthally averaged view of the rotating convection paradigm by Montgomery and Smith (2014) (see their Fig. 13d). Our focus is exclusively on the transitioning phase where the storm intensifies from a tropical storm to become a mature hurricane or typhoon. Figure 2.1 shows the conceptual model for tropical cyclone intensification. In this model there are five processes supporting the intensification process:

(1) Boundary layer inflow and spin up of ambient air mass;

(2) Evaporation over the sea;

(3) Rising of air in the eyewall;

(4) Latent heat release in the eyewall;

(5) Gradient wind adjustment to the modified temperature field.

These five reinforce each other in the intensification process as suggested by the results of a previous study based on the Ooyama-model (Frisius and Lee 2016).

This mechanism should hold regardless of if the cyclone is truly axisymmetric or not.[1] As air flows

[1] Montgomery and Smith (2014) found in a three-dimensional model that eddy dynamics contribute to the formation of the eyewall cloud. On the other hand, Persing et al. (2013) showed that the intensification mechanism also works in axisymmetric models though the evolution is less coherent in their specific axisymmetric model simulations.

into the cyclone over the warm ocean waters it carries high angular momentum inward although some amount is lost by surface drag. Through the process of evaporation it increases enthalpy and CAPE is generated.[2] This high enthalpy is transported to the eyewall and overshoots the radius of maximum gradient wind (RMGW). Due to the deacceleration by the overshoot the air flows upward into the free troposphere where convection consumes the CAPE. The unbalanced boundary layer dynamics leading to the overshoot turns out to be important for generating an updraught inside of the RMGW (see (Frisius and Lee 2016)). The convective activity releases enough latent heat to compensate adiabatic cooling because of the increase of boundary layer entropy with time and some existing CAPE. The negative radial gradient of warming at the outward side of the updraught induces a spin-up of the tangential wind by the gradient wind adjustment mechanism that can be well described by the Sawyer-Eliassen equation based on the balance approximation. Indeed, Shapiro and Willoughby (1982) showed with this equation that heating inside of the wind maximum leads to inflow and angular momentum import. Furthermore, Frisius and Lee (2016) found that in the Ooyama model the latent heat release is the most important process for the spin-up of the tangential wind above the boundary layer. The increase of gradient wind causes in turn a stronger boundary layer inflow, more angular momentum import and more evaporation from the sea surface. A positive feedback loop results.

As noted by Frisius and Schönemann (2012) there can be a negative vertical entropy gradient in the boundary layer due to the fact that the boundary layer air experiences an increase of entropy due to surface fluxes. Such a negative entropy gradient creates a convectively unstable environment thus having CAPE. On the other hand, convection acts to reduce the thermodynamic disequilibrium that exists between the troposphere and the boundary layer. Therefore, surface fluxes of heat and moisture are necessary for maintaining the CAPE.

In the following we explain the possible impact of CAPE on intensification. The tropical cyclone's most intense winds are located beneath the eyewall. In the eyewall air flows approximately along slantwise surfaces of constant angular momentum (Willoughby 1988). This yields a flow that is neutral to symmetric instability if the up- and outward flow is fast enough to nearly conserve its specific entropy. We can take full advantage of this fact by replacing the radius $r$ with a new radial

[2] We note that CAPE is not a fluid property, i.e., it does not exclusively depend on the properties of the specific fluid parcel. Therefore, generation refers to the area integrated CAPE.

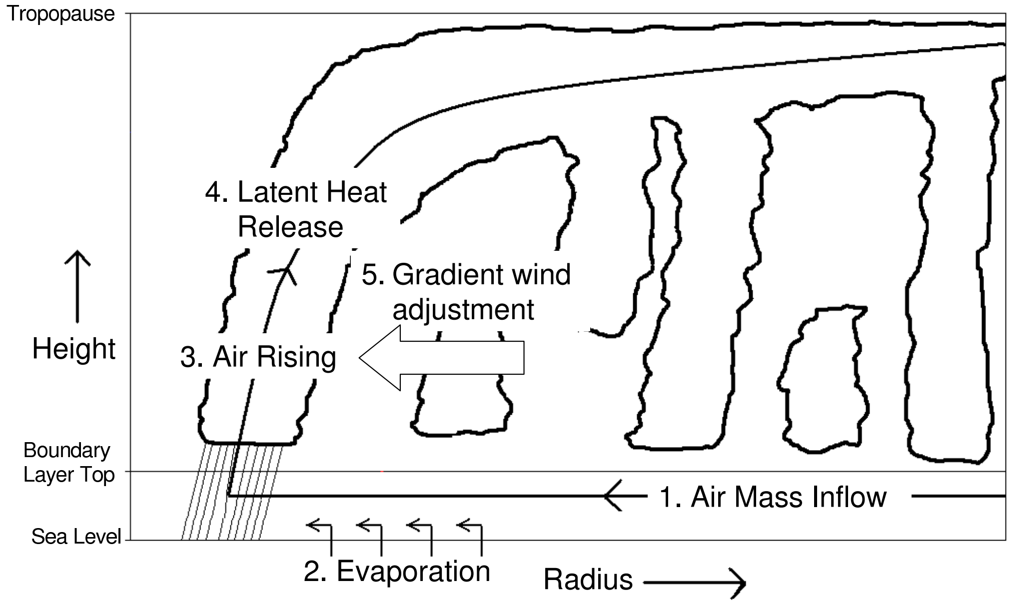



Figure 2.1: Conceptual model of intensification. Figure taken from Lee and Frisius (2018).

coordinate, namely the potential radius $R$ introduced by Schubert and Hack (1983). It is given by

$$R = \left(\frac{2}{f}vr + r^2\right)^{\frac{1}{2}} = \left(\frac{2}{f}M\right)^{\frac{1}{2}} \ , \tag{2.1}$$

where $v$ denotes the tangential wind, $f$ the Coriolis-parameter and $M = vr + fr^2/2$ the angular momentum. Entropy surfaces run in the neutral case vertically in potential radius space and, therefore, entropy becomes a vertically independent variable. As noted by several authors such as LaSeur and Hawkins (1963), the circulation in a tropical cyclone is observed to be approximately in gradient wind balance. The saturation entropy gradient in the atmosphere is related to the gradient wind through the thermal wind balance equation (Emanuel 1986). Using the thermal wind balance equation, Frisius and Schönemann (2012) found in potential radius space the approximate relation:

$$\frac{v_b^2}{R} = -\frac{1}{2}\left(T_s - T_{out}\right)\frac{\partial s_b}{\partial R} + \frac{\partial E_{SC}}{\partial R} \ , \tag{2.2}$$

where $v_b$ is the tangential wind at the top of the boundary layer, $T_s$ is the sea surface temperature, $T_{out}$ is the outflow temperature, $s_b$ the specific boundary layer entropy and $E_{SC}$ is slantwise CAPE (SCAPE). The first term on the right hand side of Eq. (2) yields the thermal wind when the stratification is neutral to slantwise convection. This assumption has been used in the potential intensity theory by Emanuel (1986) and in the simplified WISHE model by Emanuel (1997). Obviously, further intensity can result when SCAPE increases in radial direction. Therefore, heating in the eyewall induces a larger intensity increase when SCAPE is present since the heating reduces SCAPE of the eyewall and increases the radial SCAPE gradient. This corresponds to a negative radial temperature gradient which is correlated with the wind intensity at the surface. Large initial SCAPE has a high potential to create a significant radial increase of SCAPE in the intensification phase but a requisite is that the release of SCAPE occurs predominantly in the inner core of the cyclone where the frictionally induced inflow converges. The effect of heating can be clearly seen in one of the model experiments performed by Montgomery et al. (2009) (see their Fig.5). In this run the equivalent potential temperature above the boundary layer rises inside of the radius of maximum winds (RMW) while outside of this radius there is only a little increase. This results in a positive gradient of SCAPE and a tangential wind increase by the gradient wind adjustment. According to WISHE theory no SCAPE is present during intensification. Therefore, intensity builds entirely on the radial decrease of boundary layer entropy $s_b$. However, there is no clear mechanism why it should do so. The surface enthalpy fluxes increase $s_b$ at the radius of maximum winds. Then, the radial gradient of $s_b$ becomes positive in the inner core of the cyclone so that Eq. (2) does not yield a solution. Indeed, Fig.5 of Montgomery et al. (2009) reveals in the intensification phase no enhanced negative gradient of equivalent potential temperature in the boundary layer at the RMW which cannot be true if WISHE theory is correct. Furthermore, it can be shown that heating by wind speed dependent surface fluxes lead to outward migration of the tangential wind maximum (see (Frisius 2006)). Emanuel (1997), Gray and Craig (1998) and Frisius (2006) circumvented this issue by introducing a factor taking into account that convection at larger radii mainly act to moisten and not to heat the atmosphere because of a dryer atmosphere. Montgomery et al. (2009) doubted the physical justification of this factor and interpret it as an ad-hoc parameter. Later, Emanuel (2012) modified his theory by taking into account that the outflow temperature $T_{out}$ is not a constant but depends on the outflow stratification. With the assumption of a constant Richardson number in the outflow he was able to describe intensification by WISHE without a correction factor. However, Persing et al. (2013) found on the basis of CM1 model simulations that the predictions of the revised theory do not hold. These findings suggests that vanishing CAPE (or SCAPE) is too strong a model constraint and that CAPE indeed represents an important key factor for tropical cyclone intensification."

# Chapter 3

# Low-Order Box Model

## 1. Introduction

Simple models simplify complex processes making the understanding of the dynamics of a very complicated process much easier. Gray and Craig (1998) argues that these simple models are useful in tropical cyclone (TC) research because these type of models focus on the more important processes that govern intensification and as a result can provide potential experiments that can be conducted to test how certain environmental factors can affect the growth of these cyclones. Due to the complexity of TC intensification and to the fact that there are still many unknowns in the dynamics and thermodynamics, a simple box model provides a great tool to conduct experiments to better understand the role CAPE plays in intensification. In this chapter we investigate the mechanism of intensification by using a low-order box model. Transports of entropy in the boundary layer to the eyewall via the secondary circulation is the key to understanding why CAPE is critical to the cyclone's survival. We also take note that certain climatological factors which affect the stability of the atmosphere can impact on the development of a vortex, resulting in certain situations where there is no TC formation. While retaining the important processes that are involved in TC intensification, the simplicity of the model provides a less complicated approach to an area of TC research that has become of great importance. Our results from this model, also bring into question the validity of the WISHE paradigm that has gained much popularity and recognition. As a consequence of this, the model is able to explain why WISHE may need modifications to it so that it can fully capture all the relevant processes that dominate during intensification. The WISHE intensification theory is widely cited as the mechanism for TC intensification in several textbooks (e.g Holton (2004)). WISHE supports a positive feedback system between the surface airflow and the evaporation of water over the warm ocean waters into an atmosphere that is in thermodynamic disequilibrium due to the strong entropy gradient that exists between the warm ocean waters and the air. The experiments we conduct highlight the weakness of WISHE theory and it provides an explanation for it. The low-order box model was developed by (Schönemann and Frisius 2012) with the belief that TCs are independent dynamical systems whose state can be described by certain external parameters. Schönemann and Frisius (2012) used the model to study TCs' steady state behaviour along with the initial growth of TCs in the framework of CISK and WISHE. Due to its simplicity, we use the model to understand intensification in an environment that made a very complex subject very simple and easy to understand.

# 2. Model Description

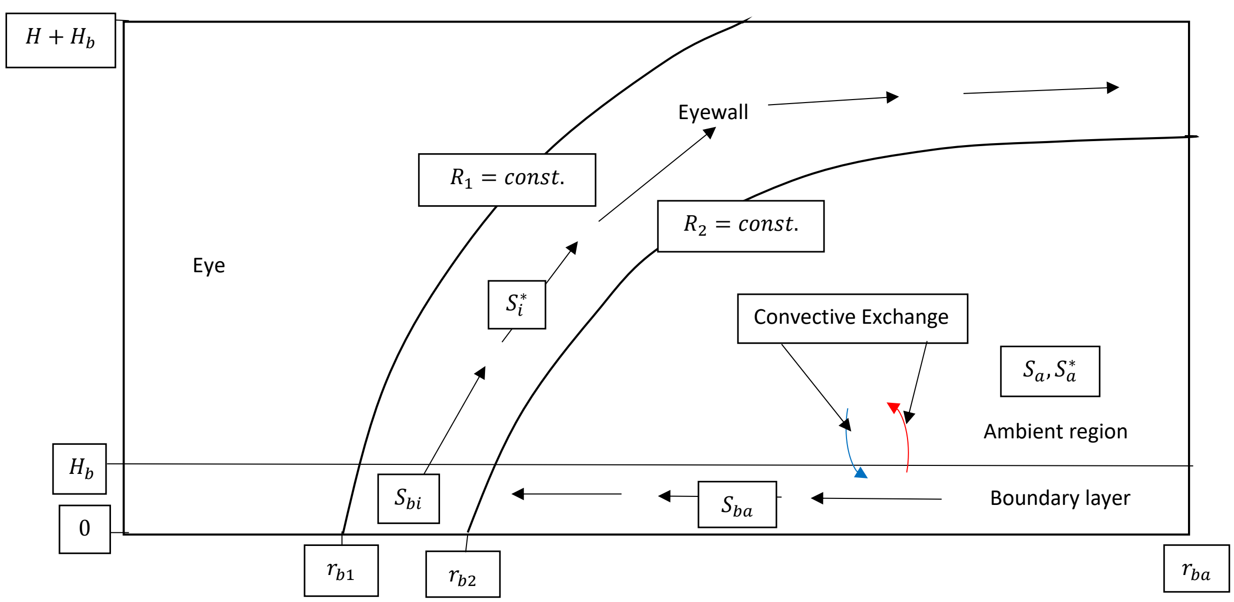



Figure 3.1: Sketch of the low-order tropical cyclone model where s is specific entropy, $R$ is the potential radius, r the physical radius and the indicies b,i and a stand for boundary layer, eyewall and ambient region, respectively.

The model is formulated in cylindrical coordinates with the assumption that the vortex is axisymmetric with the air flow being in gradient wind balance. Figure 3.1 shows a sketch of the low order model where the black arrows show the direction of the large-scale secondary circulation. The arrows in the horizontal plane represent the boundary layer inflow while the arrows in the quasi-vertical represent the outflow above the boundary layer. The air flows in the boundary layer, up through the eyewall until it reaches tropopause where it flows out to some distance. The red arrow represents convective updrafts - importation of high entropy air into the free atmosphere - while the blue arrow represents convective downdrafts - the importation of low entropy air into the boundary layer. Both those arrows represent convective exchange in the ambient region - outside the eyewall. The model variable is specific entropy $s$. The boundary layer and the free atmosphere form the two layers of the model while isopleths of angular momentum divide the model into three regions -the eye, the eyewall and the ambient region- where all three regions have partitions into the the free atmosphere and the boundary layer. These isopleths mark the boundaries of the both the inner ($r_{b1}$) and outer ($r_{b2}$) radii resulting in the eyewall being enclosed by these angular momentum surfaces. These physical radii do move in the course of the development of the cyclone. Additionally, the potential radii are located at these angular momentum surfaces for the inner ($R_1$) and outer($R_2$) radii where they are constant in time. Therefore, the regions form rigid boxes in potential radius space. Potential radius results from a coordinate transformation introduced by Schubert and Hack (1983) where the physical radius becomes a dependent variable. Potential radius $R$ is the radius where a particle must move to obtain zero tangential velocity while conserving its angular momentum. This coordinate system is ideal to describe slantwise convection such as it takes place in the eyewall of tropical cyclones.

The boundary layer height $H_b$ is set to be constant. The slab boundary layer model of Schubert and Hack (1983) makes the airflow of the boundary layer a function of the balanced tangential wind at the top of the boundary layer. Limitations of this type of models have been noted by Smith and Montgomery (2008) and by Frisius and Lee (2016), however, due to the simplicity of the model dynamics it is used in this low-order model. The eye develops passively so there are no exchanges of energy and mass with the eyewall, and therefore no equations for the eye. Within the eyewall the assumption is made that angular momentum and mass are conserved and movement of the eyewall during development is due to gradient wind adjustment. We make the assumption that all the heating takes place within the radius of maximum gradient wind (RMGW). To preserve the simplicity of the model, the maximum wind speed is equated to the maximum gradient wind speed. There is no ascent in the ambient region and the only calculation that is needed in this region is the boundary layer mass flux from the ambient region into the eyewall region.

The saturation entropy of the eyewall $s_i^*$ is modified by diabatic processes and the mass flux from the boundary layer. The specific entropy $s_{bi}$ is modified by surface heat fluxes and inflow from the ambient region. The specific boundary layer entropy $s_{ba}$ is modified by surface fluxes, radial advection, convective cooling and downwelling of low-entropy air from the free atmosphere. The value of the entrainment parameter, $\delta$, which describes the entrainment of air into the ambient boundary layer lies between 0 and 1. The eyewall is assumed to be neutral to slantwise convection in potential radius space. Entropy of the ambient region $s_a$ is measured as an anomaly of entropy in the far-field environment. The governing equations of the model are

$$\frac{ds_i^*}{dt} = \left(\Psi_{b2} + \frac{M_{bi}}{\tau_C}\right)\frac{s_{bi} - s_i^*}{M_i} + \frac{s_a^* - s_i^*}{\tau_E}, \tag{3.1}$$

$$\frac{ds_{bi}}{dt} = \Psi_{b2}\frac{s_{ba} - s_{bi}}{M_{bi}} + \frac{s_i^* - s_{bi}}{\tau_C} + \frac{C_H}{2H_b}\left(|v_{b1}| + |v_{b2}|\right)\left(s_{oi} - s_{bi}\right), \tag{3.2}$$

$$\frac{ds_{ba}}{dt} = \Psi_{b2}\frac{\delta s_a - s_{ba}}{M_{ba}} + \frac{C_H}{2H_b}|V_{b2}|\left(s_{oa} - s_{ba}\right) + \frac{s_a - s_{ba}}{\tau_C}\mathcal{H}\left(s_{ba} - s_a^*\right), \tag{3.3}$$

where the indices $b$, $i$ and $a$ stand for boundary layer, eyewall and ambient region, respectively. The asterisk ($^*$) signifies that the variable is being evaluated by assuming that there is water vapour saturation. $\mathcal{H}(\cdots)$ is the Heaviside function, $M$ is the mass of the corresponding region,$\Psi_{b2}$ is the radial mass flux of the boundary layer at $r_{b2}$, $C_H$ is the surface transfer coefficient for enthalpy, $s_o$ is the sea surface entropy, $\delta$ is the entrainment parameter, $\tau_C$ is the time scale for convection, $\tau_E$ is the timescale for Newtonian cooling which represents a simple parameterisation for radiation in the model, $v_{b1}$ is the gradient wind at the inner eyewall boundary and $v_{b2}$ is the gradient wind at the outer eyewall boundary.

We have modified the original model formulation by including convective exchange in the eyewall. In the ambient region convective exchange only takes place where $s_{bi} > s_a^*$. This allows the vortex to intensify under the WISHE paradigm. The basis of WISHE is that convection acts so fast that the atmosphere is always kept in a moist neutral state. As a result intensification is possible for very small $\tau_C$.

The model dynamics restricts to the WISHE intensification when $\tau_C \to 0$. As a consequence $s_i^* = s_{bi}$ and $s_{ba} = s_a^*$. This causes the governing equations to reduce to one equation and the low-order model now becomes:

$$\frac{ds_i^*}{dt} = \Psi_{b2}\frac{s_a^* - s_i^*}{M_i + M_{bi}} + \frac{s_a^* - s_i^*}{\tau_E}\frac{M_i}{M_i + M_{bi}} + \frac{C_H}{2H_b}\left(|v_{b1}| + |v_{b2}|\right)\left(s_{oi} - s_i^*\right)\frac{M_{bi}}{M_i + M_{bi}}. \tag{3.4}$$

Since the surface fluxes heat the eyewall immediately this equation can explain the intensification is by WISHE at the beginning. This equation describes the adaptation of the eyewall to the sea surface entropy $s_{oi}$ where the timescale for this process is

$$\tau_W = \frac{2H_b}{C_H\left(|v_{b1}| + |v_{b2}|\right)}\frac{M_i + M_{bi}}{M_{bi}}. \tag{3.5}$$

Eventually the secondary circulation entrains low entropy air from the environment,($s_a^*$) into the eyewall which retards intensification, and this process takes place with the timescale

$$\tau_S = \frac{M_i + M_{bi}}{\Psi_{b2}}. \tag{3.6}$$

Figure 3.2 shows the times scales for WISHE, $\tau_W$ and of the secondary circulation, $\tau_S$ as a function of the tangential wind $v_{b2}$. From the figure we see that the WISHE mechanism is present for low values of $v_{b2}$. When $v_{b2}$ takes on values of that are typical of a tropical cyclone in the intensification phase the timescale, $\tau_S$ becomes smaller than that of $\tau_W$. So we can infer that the negative impact of secondary circulation develops before the WISHE mechanism can substantially act on the TC. This has important implications for the energy production and maintenance of the storm. The entropy fluxes of the secondary circulation are the ones that dominate the eyewall entropy budget.

This has already been proven by Wang and Xu (2010) where they noted that the surface entropy fluxes outside the eyewall in the boundary layer contribute significantly to the energy balance in the eyewall. They came to this conclusion after evaluating the balance hypothesis by Emanuel (1997, 1995)- energy production is equal to frictional dissipation at the RMW - and noting that the frictional dissipation rate was larger than the energy production rate under the eyewall while the energy production rate was greater than the frictional dissipation rate outside the eyewall. Therefore, to achieve balance a portion of the energy produced outside will have to offset the excess loss due to frictional dissipation. Wang and Xu (2010) further highlight the fact that Emanuel (1997) neglects the activities outside the eyewall. Emanuel (1997) only considered energy production and dissipative heating near the RMW in the eyewall region. To account for all these processes that occur outside the eyewall, Emanuel (1997) introduced a new parameter $\beta$ which crudely accounts for any process that affects the distribution of entropy outside the eyewall such as convection and large-scale downdrafts in the boundary layer. The effect of using this parameter is that it reduces entropy in the boundary layer.

# 3. Experimental Design

To execute our investigation and determine the mechanism of intensification we conduct five experiments. To ensure cyclogenesis, relative humidity of the boundary layer and the free atmosphere is set to 80%, the sea surface temperature to 28°C and $\tau_C$ = 10 hrs unless it is stated otherwise. All other model parameters are listed in table 3.1.

- The reference experiment REF, assigns a finite value of 0.6J/kg/K for $s_i^*$ and $s_{bi}$ while setting $s_a^*$ and $s_{ba}$ equal to zero.
- The HIGHCAPE experiment includes initial environmental CAPE by reducing the temperature of the ambient region $T_a$ by 2.5K.
- The STABLE experiment increases the temperature $T_a$ by 2.5K.
- The WISHE experiment integrates equation 3.4 so any CAPE is suppressed.
- The NOEXCHANGE experiment assigns an infinite timescale to $\tau_C$ so there is no convective

exchange in the ambient region.

Table 3.1: Parameter values for the low order tropical cyclone model experiment REF.

| Notation | Value | Meaning |
|---|---|---|
| $r_{ba}$ | 420 km | Outer radius |
| $\tau_E$ | 48 h | Time scale for Newtonian cooling |
| $\tau_C$ | 10 h | Time scale for convection |
| $C_H$ | 0.003 | Surface transfer coefficient for enthalpy |
| $C_D$ | 0.003 | Drag coefficient |
| $H$ | 13.5 km | Height of free atmosphere layer |
| $H_b$ | 1.5 km | Boundary layer height |
| $f$ | 5x10$^{-5}$s$^{-1}$ | Coriolis parameter |
| $\delta$ | 0.25 | Entrainment parameter |
| $R_1$ | 90 km | Inner potential radius of eyewall |
| $R_2$ | 180 km | Outer potential radius of eyewall |
| $T_t$ | 203.15 K | Tropopause temperature |
| $T_s$ | 301.15 K | Sea surface temperature |
| $h_a$ | 80% | Relative humidity, ambient region |
| $h_b$ | 80% | Relative humidity, boundary layer |

# 4. Results

Figure 3.3 shows the time evolution of the surface pressure, the tangential wind ($v_{b2}$) and the radius of maximum wind (RMW) for the various experiments. This model has no gestation period and as a result development starts immediately. In models where there is an initialisation or gestation period, there is a characteristic weakening or decaying of the vortex after initialisation which lasts only for a few hours and then it intensifies. Both Reed and Jablonowski (2011) and Smith et al. (2009) have explained that the characteristic weakening is due to surface friction. Reed and Jablonowski (2011) pointed out that the lack of the secondary circulation also contributes to the

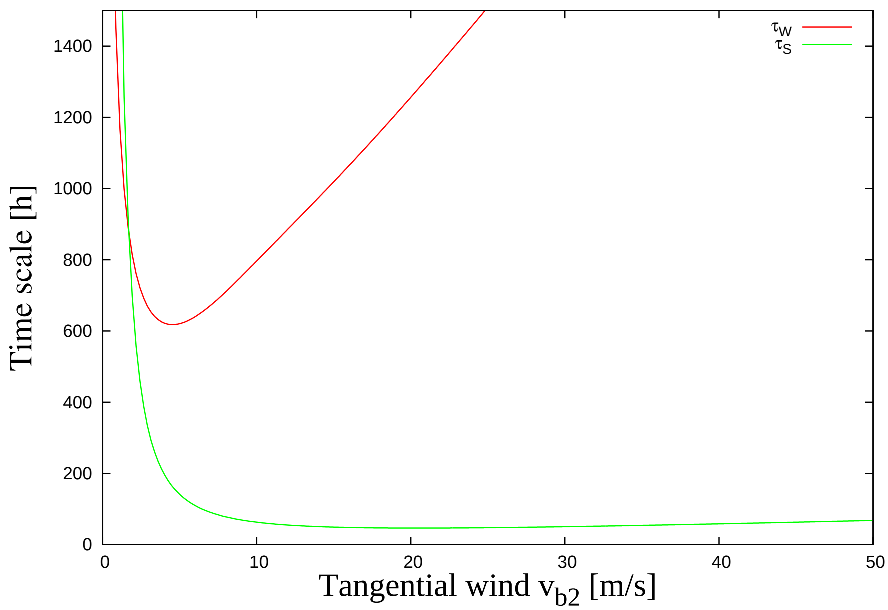



Figure 3.2: Time scale of WISHE $\tau_W$ (red curve) and of the secondary circulation $\tau_S$ (green curve) as a function of tangential wind $v_{b2}$.

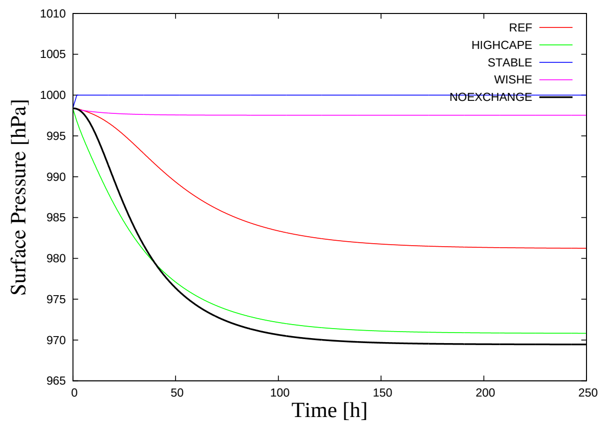
vortex decaying after initialisation. In this model the eyewall region is saturated outright from the beginning so the TC can intensify immediately, as a result there is no decaying of the vortex.

All experiments start with a surface pressure of 998 hPa. There is a pressure drop in all except in the WISHE and the STABLE experiments as seen in figure 3.3a. The HIGHCAPE and NOEXCHANGE experiments have a steep decline where both have a reduction in pressure at the same rate. Around 100 hours they both stabilise. The surface pressure drop correlates with the tangential wind increase we see in both these experiments in figure 3.3b. Their $v_{b2}$ like all the other experiments has an initial value of 11 m/s. There is a steep rise in tangential wind where HIGHCAPE intensifies a little bit faster than NOEXCHANGE. The final $v_{b2}$ of HIGHCAPE is 65m/s which is substantially higher than the REF experiment with a final $v_{b2}$ of 51m/s. This contradicts what Persing and Montgomery (2005) have stated unequivocally, that environmental CAPE has no impact on modeled hurricane intensity. Our results show a clear indicator that there is a difference in excess of 10m/s in intensity. The tropical cyclone is not insensitive to CAPE in the simple model.

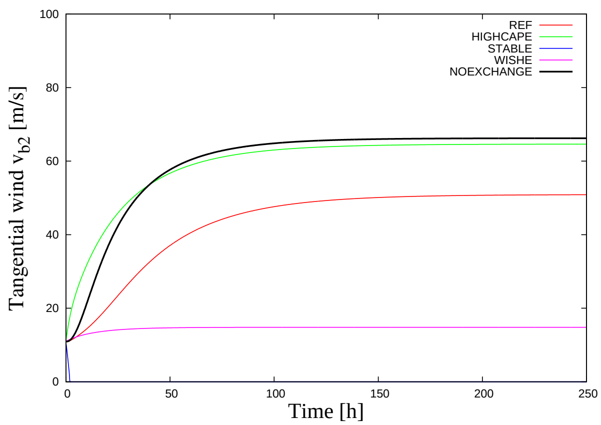
The WISHE experiment shows nearly no intensification. It starts with an initial value of 11m/s like all the other experiments and it rises to about 15m/s and stays there throughout the time period. This result is expected because the frictionally induced secondary circulation transports low entropy air into the eyewall. This hinders the development. Looking at figure 3.2 we can see that the time it takes for the secondary circulation to develop is much less than the time it takes for WISHE mechanism to act on the vortex. This explains why the WISHE experiment failed to develop a hurricane-type vortex. Stabilising the atmosphere, clearly provides conditions that do not support the development of a cyclone. In the STABLE experiments in figure 3.3 we see that there is a rapid decay of the vortex. The ambient region is warmer than the cyclone core but a positive temperature difference needs to exist between the cyclone and the ambient region. Factors that affect this will impact on the ability of the cyclone to intensify.

Figure 3.3c shows the time evolution of the RMW, where since $v_{b2}$ is equal to the maximum gradient wind, $r_{b2}$ can be taken to be the radius of maximum wind. There are many ways one can define the size of a tropical cyclone. Carrasco et al. (2014) uses RMW as a measure of the size of a tropical cyclone and we do it likewise. From the figure you can see that the vortex contracts when it intensifies for the three shown experiments REF, HIGHCAPE and NOEXCHANGE. As expected, the HIGHCAPE and NOEXCHANGE have a much steeper decline in size of the vortex. Both HIGHCAPE and NOEXCHANGE vortices maintain roughly the same size since both their radii end at 13km.

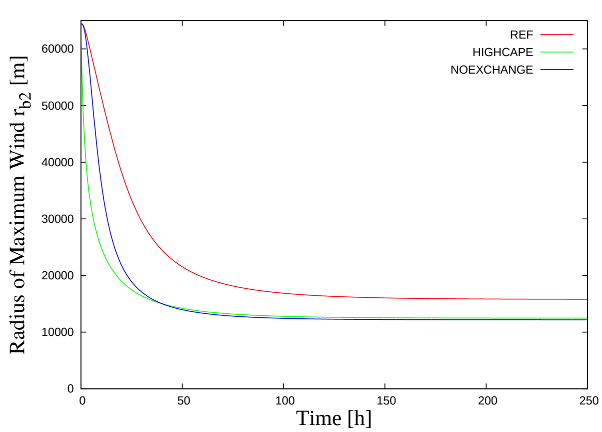
Figure 3.4 gives the same information as that of figure 3.3 but it shows results of experiments STABILISE and DESTABILISE in which we stabilise and destabilise, respectively the atmosphere by modifying the boundary layer. In experiment STABILISE we use a value of -10 for $s_{ba}$ and $s_{bi}$, and in experiment DESTABILISE we use a value of 10 for $s_{ba}$ and $s_{bi}$. This is to test to see if there is any impact the way CAPE is introduced may have on the intensification. Is the cyclone sensitive to how CAPE is introduced? To stabilise the atmosphere via boundary layer modification of the relative humidity would require drying out the boundary layer, and conversely to destablise would be to add moisture to the layer. In the model experiment REF the relative humidity parameter in the boundary layer and in the free atmosphere are both set at 80%. To ensure we maintained the necessary stability requirements, adjustments were made to the boundary layer entropies ($s_{ba}$ and $s_{bi}$). Once the model begins, the ambient air re-establishes convective neutrality by decreasing the humidity of the boundary layer.




(a) Minimum surface pressure for all experiments.




(b) Tangential wind $v_{b2}$ for all experiments.



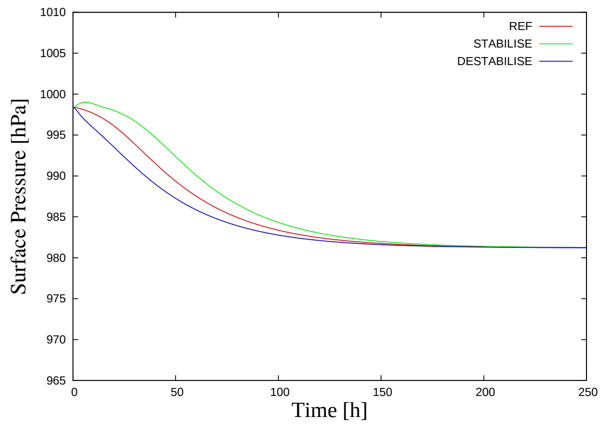

(c) Radius of Maximum Wind $r_{b2}$ for REF, HIGHCAPE and NOEXCHANGE experiments.

Figure 3.3: Time evolution of the various experiments.

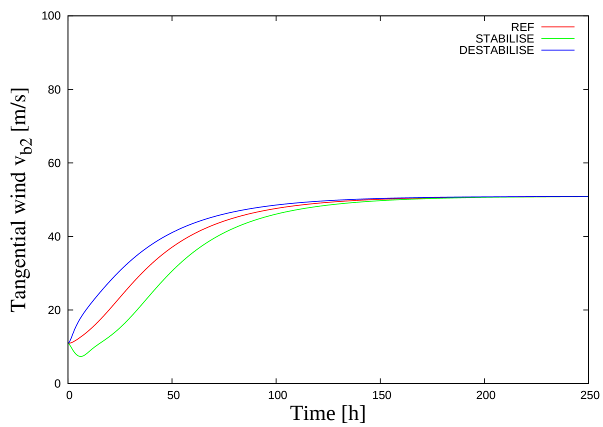
All the experiments have the same initial $v_{b2}$ and surface pressure as those of the previous experiments. Not surprising the DESTABILISE experiment has the fastest intensification evidenced by the steepest drop in surface pressure and the associated increase in $v_{b2}$. Unlike the other experiments, the STABILISE experiment reveals that the vortex weakens before it intensifies. The increase in surface pressure accompanied by the associated decrease $v_{b2}$ lasts for about 8hrs and then it begins to intensify. This may be due to the drying out of the boundary layer by using lower values for $s_{ba}$ and $s_{bi}$. In fact, the STABILISE experiment had the slowest onset of intensification. However, even though all three experiments had a different intensification rate they all coincided with the REF experiment around 175 hrs, ending with a value of 51m/s. The size of all the vortices started at 65km and all dropped to the same size as the REF to be 17km.

These experiments reveal a significant fact about CAPE. When you compare these results with those of the previous experiments, it is evident that the manner in which CAPE is added does impact the intensification rate. The intensification phase of both experiments, STABILISE and DESTABILISE, are significantly less strong than of the HIGHCAPE and NOEXCHANGE experiments. There is a clear distinction in how the cyclone is intensifying depending on how CAPE was included. In the latter group of experiments, CAPE is included by modifying the boundary layer. A shallow layer of air where the entropy can be modified very quickly due to surface fluxes and convective downdrafts. Surface fluxes generate CAPE and increase the boundary layer entropy while convective downdrafts transport low-entropy air from the free atmosphere into the boundary layer thus lowering the boundary layer entropy. When CAPE is included by cooling the deep free atmospheric layer (as in the first group of experiments), convection acts very slow to remove the instability while the surfaces fluxes are unable to remove the stability in a deep stable layer. As a consequence CAPE included from modification to the deep free atmosphere has a much greater impact than from if it were only included in the shallow boundary layer.

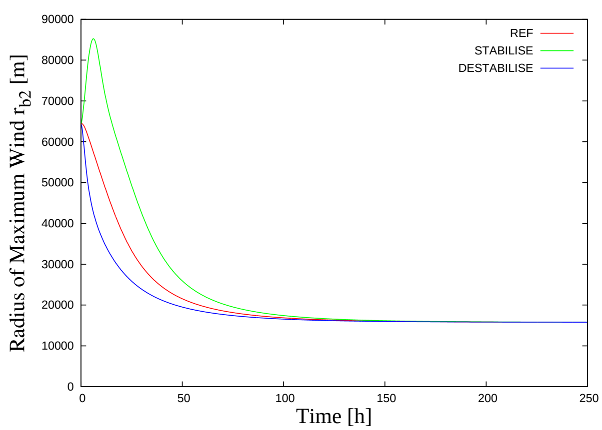
To further understand the influence CAPE has on intensification we can analyse its time evolution throughout each experiment. Since the flow within the eyewall is not vertical but slantwise we calculate slantwise convective available potential energy (SCAPE) instead of CAPE. SCAPE of the eyewall ($E_{SC,i}$) and the SCAPE of the ambient region ($E_{SC,a}$) as per Emanuel (1994) is calculated




(a) Minimum surface pressure for all experiments.




(b) Tangential wind $v_{b2}$ for all experiments.




(c) Radius of Maximum Wind $r_{b2}$ for REF,STABILISE and DESTABILISE experiments.

Figure 3.4: As in Fig 3.3 but the results of experiments REF, STABILISE and DESTABILISE are shown.

as follows:

$$E_{SC,i} = (T_s - T_t)(s_{bi} - s_i^*), \tag{3.7}$$

$$E_{SC,a} = (T_s - T_t)(s_{ba} - s_a^*). \tag{3.8}$$

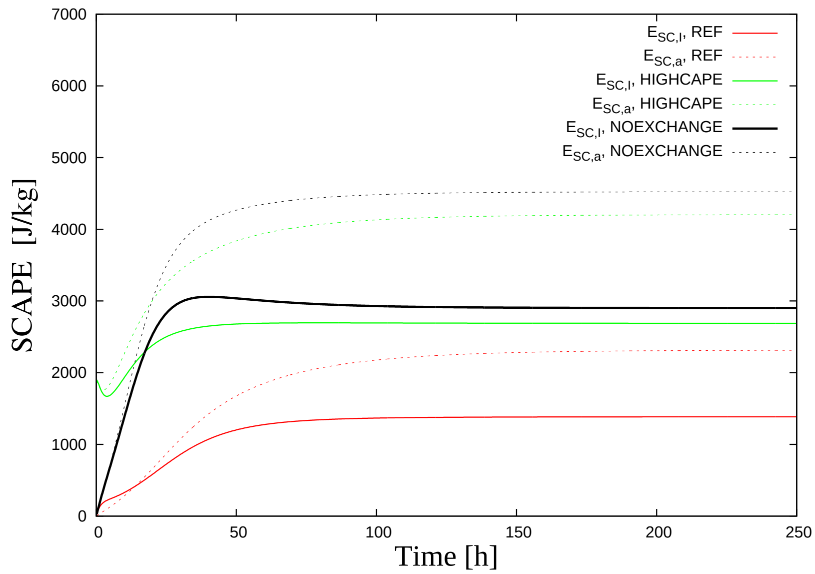



Figure 3.5: $E_{SC,i}$,(solid curves) and $E_{SC,a}$(dashed curves) as a function of time for the experiment REF (red curves). HIGHCAPE (green curves) and NOEXCHANGE (black curves).

Figure 3.5 shows the evolution of SCAPE for the different experiments in the eyewall and in the ambient region. Scape is zero at the beginning for REF and NOEXCHANGE experiments which is expected, but it starts off at 1900 J/kg for the HIGHCAPE experiment. As the vortex intensifies the amount of SCAPE increases steadily in all experiments but at different rates. We also notice that the SCAPE in the ambient region for all experiments are significantly higher than the SCAPE in the eyewall region. Experiments HIGHCAPE and NOEXCHANGE have a larger difference in their ambient region SCAPE and eyewall SCAPE. It is no surprise that the eyewall's SCAPE for all experiments is significantly less than what is in the ambient region. Within the eyewall convective processes consume SCAPE during the intensification phase which ends around 50hrs for HIGHCAPE and NOEXCHANGE experiments, and at around 75hrs for the REF experiment,then it starts to enter it's mature steady-state phase as seen in figure 3.3. This corresponds to Figure 3.5 REF's eyewall SCAPE stabilising around 60hrs, around 30 hrs for SCAPE in the eyewall of

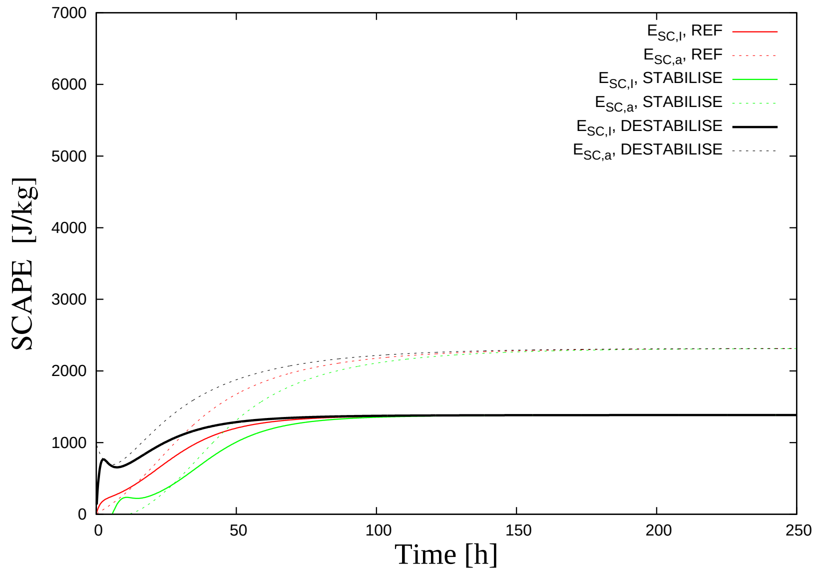



Figure 3.6: $E_{SC,i}$,(solid curves) and $E_{SC,a}$(dashed curves) as a function of time for the experiment REF (red curves). STABILISE (green curves) and DESTABILISE (black curves).

NOEXCHANGE and HIGHCAPE experiments. The time for SCAPE to remain constant will be less than the time it takes for the vortex to intensify.

Figure 3.6 shows the same thing as figure 3.5 but for the experiments where the boundary layer was modified. At the beginning, all eyewall CAPE values are zero initially but their developments take a different trajectory until they coincide with the REF experiment as expected. Both the STABILISE and DESTABILISE experiments have a minor increase around 25hrs and 5hrs respectively in the eyewall, which does not last for long and then they drop back down and have a steady increase until around 75hrs for both where SCAPE remains constant in the eyewall. The ambient region SCAPE is larger than the eyewall but its difference is significantly much less than the difference for the experiment HIGHCAPE where the free atmosphere temperature was decreased. SCAPE being increasingly higher in the ambient region to the eyewall, serves to enhance the gradient wind and thus increase the intensification rate.

# 5. Summary

We used a simple low order box model to investigate the intensification mechanism for TC development. Through a series of experiments showing time evolution of each vortex in the conceptual model, we were able to provide a possible mechanism of intensification enhancement due to the presence of CAPE. Five experiments were conducted where in the experiment STABLE the temperature of the ambient air was increased by 2.5K creating a stable atmosphere, in the experiment HIGHCAPE the ambient air was decreased by 2.5K creating an environment with high CAPE, in the experiment WISHE we applied the WISHE mechanism by creating a convective neutral environment by setting $\tau_C$ to zero so that the boundary layer entropy of the ambient region was made equal to that of the free atmosphere and similarly the eyewall boundary layer entropy was made equal to that of the entropy in the eyewall in the free atmosphere. In the experiment NOEXCHANGE we applied an infinite timescale for $\tau_C$ which allowed no convective exchange to take place in the ambient region, and finally the experiment REF was the reference run which was used as our base for comparison.

The results of the experiments prove that the presence of CAPE is an essential part of intensification, and additionally, the WISHE mechanism does not support the development of a vortex. Both HIGHCAPE and NOEXCHANGE experiments exhibited a similar pattern in their intensification rates. HIGHCAPE may have had an onset of intensification first and quickly but by around 40 hours both were stabilising close to each other with NOEXCHANGE having a slightly higher final intensity. As a result, convective activity with a finite timescale does have a damping effect on intensification and on the final intensity. The WISHE mechanism stresses the point that convection happens so fast that the atmosphere is neutral to convective instability. These results prove otherwise. Additionally, since convective activity in the ambient region has a damping effect on intensification as evidenced by comparing NOEXCHANGE with REF we are provided with another reason why the WISHE mechanism for intensification fails. If convection were to happen at a very small finite timescale the vortex would not have been able to attain hurricane force winds which is what was the case for the WISHE experiment. This reinforces the fact that the WISHE paradigm for intensification cannot fully account for the thermodynamical processes that control the intensification phase of a TC.

We further tested to see if there is a significance in the way CAPE is introduced by making modifications to our boundary layer entropy values which affected the relative humidity values thus creating a stable or unstable environment. Including CAPE in a deep layer of the atmosphere as opposed to a very shallow thin layer makes a noted significance. The shallow boundary layer is affected by surface fluxes and convective exchange processes that both affect the generation and depletion of CAPE. These process occur at very short time scales, so modifying the boundary layer alone to include CAPE did initially enhance intensification for a short period but as is seen in the results of the STABILISE and DESTABILISE experiments, they quickly coincided with the REF experiment proving that CAPE present in the free atmosphere has a greater impact than what is produced in the boundary layer. This fact should not overshadow the important dynamical and thermodynamical processes occurring in the boundary layer that greatly impact the intensification. It was revealed that on onset of intensification the secondary circulation takes a much shorter time to develop than for the WISHE mechanism to act on the vortex. The timescales show that at the very beginning there is some intensification due to WISHE but due to the convective fluxes bringing low entropy air into the boundary layer, the secondary circulation that has fully developed transports this low entropy air into the eyewall region thus hindering intensification. This further highlights the fact that the budget of the eyewall entropy is greatly affected by the entropy fluxes of the secondary circulation. They dominate and as such the secondary circulation, although weaker as the primary circulation, can greatly impact the strength of the vortex tangential winds and ultimately the survival of the vortex.

Our findings are in agreement with Wang and Xu (2010) where they were able to show that air flowing towards the eyewall in the boundary layer can gain a great amount of energy due to surface entropy fluxes. These fluxes outside the eyewall have an impact on the magnitude of the storm a fact that was revealed in the study. So we are shown through experiments and budget analysis in that study that transports of the secondary circulation are important to the energy balance in the eyewall of a mature TC. The WISHE-type models of Emanuel (1997), Gray and Craig (1998), Frisius (2006) and Emanuel (2012) do not have properly defined boundary layers and as such do not include the process of the transports of the secondary circulation from outside the eyewall. The experiments of Wang and Xu (2010) was in response to Emanuel (1997) who downplayed the significance of processes outside the eyewall (inner core) in the mature TC's energy balance by introducing the $\beta$ parameter which crudely accounted for processes that affect the distribution of entropy outside the eyewall such as downdrafts which import low entropy air into the boundary layer. As a result, this parameter reduces the entropy in the boundary layer and also the radial advection of it which impacts on the eyewall entropy by diminishing it. The reduction in the boundary layer entropy and the reduction in the radial advection both affect the energy balance in the eyewall and due to their competing effects it is uncertain which dominates. Other studies such as Gray and Craig (1998)and Frisius (2006) both adopted the $\beta$ parameter but they introduced it differently where the parameter represented unsaturated processes outside the eyewall that have negative effects. These unsaturated processes bring about a reduction in the heating rate outside of the eyewall.

The presence of CAPE outside the eyewall enhances the negative radial temperature gradient which causes the gradient wind to increase. This results in a stronger frictional inflow of air with higher surface heat fluxes. This high entropy air is carried to the eyewall region via the secondary circulation. It is important to note that we were unable to produce a mature TC from the WISHE mechanism. Based on these results, environmental CAPE supports intensification. When CAPE is generated by surface fluxes the secondary circulation is able to transport high entropy air from large radii outside the eyewall to the eyewall where it provides the energy needed to accelerate the tangential wind and maintain the storm from dissipation. The WISHE mechanism cannot support a hurricane vortex in this conceptual model but to further test our results we will use a model of higher complexity in the next chapter.

# Chapter 4

# Ooyama's three-layer model

## 1. Introduction

Ooyama (1969) (hereafter referred to as O69) was the first to successfully simulate a tropical cyclone to maturity. This model is an improvement of an earlier version (see Ooyama (1964)) where the cyclone never was able to reach maturity, instead the vortex kept growing. The vortex unable to attain a mature stage was already a problem within the community (see Ogura (1964)). O69 considered the circulation to be governed by an axisymmetric balanced vortex that is embedded in an atmosphere that has only three layers where the bottom layer is the boundary layer. One of the improvements was the representation of the parameter for deep convention, $\eta$, where it was viewed to be constant originally but now it can vary. As a result it can take into account the effect the warm core has on stabilising the core and the variation of equivalent potential temperature in the boundary layer. The model was built so the basic process that dominate during intensification can be understood and the response the cyclone has to changes in certain physical parameters. However, due to the approximations that are made to build the model and that the vortex is assumed to be axisymmetric, there are limitations on what can be studied. The basic outcome of O69 is that a weak disturbance grows very quickly into a mature vortex. The circulation during intensification produces a warm core where the moist static stability approaches a neutral state near the centre. Resulting in a reduction of the convective instability which impacts negatively on the intensification. The hopes of O69 was for this model to form foundation of future better models.

In this research we use a modified version of O69 which appears in Frisius and Lee (2016). In the original, the tangential winds are assumed to be in gradient wind balance but as Frisius and Lee (2016) noted this seems highly unlikely since the rate of pressure drop becomes too large to support balanced dynamics. In that study they investigated the impact using unbalanced dynamics will have on intensification. It was revealed that the most intense intensification occurred in where unbalanced dynamics were employed. Ooyama (1968, unpublished manuscript)[1] used his unbalanced dynamics in the boundary layer of his three-layer model and produced results that were more realistic than what the original model produces (see (Smith and Montgomery 2008) for more on this). O69 attracted a lot of attention (e.g., DeMaria and Schubert (1984),DeMaria and Pickle (1988), Smith (1997), Dengler and Reeder (1997)Camp and Montgomery (2001), Schecter and Dunkerton (2009), Frisius and Lee (2016)). We use it here to better understand intensification by relaxing the balanced assumption. In addition to using the balanced approximation, we added a convective parameterization scheme that does not rely on frictional convergence for convection. This allows for us to use the model to simulate the conditions that are needed for WISHE. The next sections in this chapter describes the model, explains how the experiments were conducted, show our results and a summary of everything.

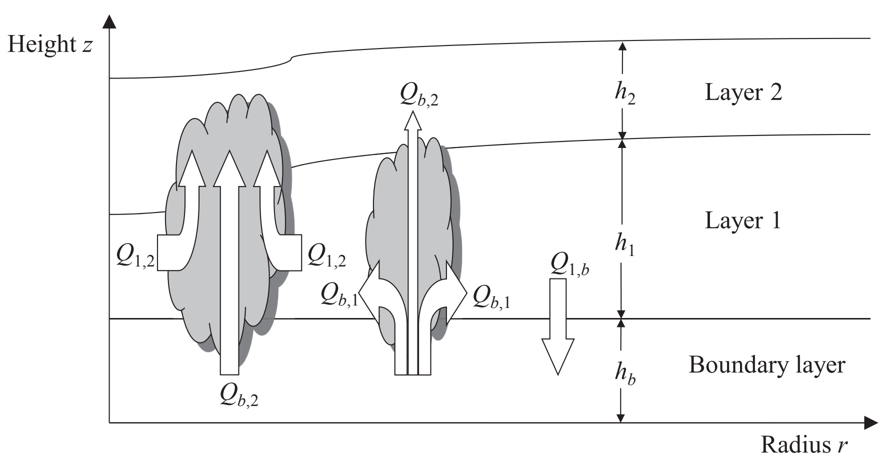



Figure 4.1: A sketch of the Ooyama model.

[1]Numerical Simulation of Tropical Cyclones with an Axisymmetric Model.

## 2. Model Description

Figure 4.1 shows a sketch of the Ooyama model. In the figure the atmosphere is divided into three layers with their respective depths of $h_b$, $h_1$ and $h_2$. There is a free interface that separates layer 1 from layer 2 but also a fixed interface that separates the boundary layer from layer 1. The densities of the boundary layer and layer 1 are identical while the density of layer 2 is smaller by a factor $\epsilon$. Convective motion moves throughout the model. This creates the mass fluxes $Q_{b,1}$, $Q_{b,2}$ and $Q_{1,2}$. $Q_{b,1}$ represents the detrainment of air from the updraught, $Q_{b,2}$ represents deep convection and $Q_{1,2}$ represents entrainment of air in the middle layers into the updraught. To ensure that mass is conserved there needs to be a downward mass flux in the boundary layer, $Q_{1,b}$. The flow is governed by the hydrostatic Boussinesq equations.

We use a modified version of the O69 three-layer model. For full details of the model please read Frisius and Lee (2016).

The governing equations for the Ooyama model are as follows:

$$\frac{\partial u_j}{\partial t} + u_j\frac{\partial u_j}{\partial r} - \left(f + \frac{v_j}{r}\right) v_j = -\frac{\partial P_j}{\partial r} + D_{v,u_j} + D_{h,u_j}, j = 1, 2 \tag{4.1}$$

This is the radial momentum equation of the free atmosphere layers.

$$\frac{\partial v_j}{\partial t} + \zeta_j u_j = D_{v,v_j} + D_{h,v_j}, j = 1, 2, \tag{4.2}$$

This is the tangential momentum equations for the free atmosphere.

$$\frac{\partial u_b}{\partial t} + u_b\frac{\partial u_b}{\partial r} - \left(f + \frac{v_b}{r}\right) v_b = -\frac{\partial P_1}{\partial r} + D_{v,u_b} + D_{h,u_b} + D_{s,u_b}, \tag{4.3}$$

This equation is radial momentum equation for the boundary layer.

$$\frac{\partial v_b}{\partial t} + \zeta_b u_b = D_{v,v_b} + D_{h,v_b} + D_{s,v_b}, \tag{4.4}$$

This is the tangential momentum equation of the boundary layer.

$$\frac{\partial \theta_{e,b}}{\partial t} + u_b \frac{\partial \theta_{e,b}}{\partial r} = D_{v,\theta_{e,b}} + D_{h,\theta_{e,b}} + D_{s,\theta_{e,b}}, \tag{4.5}$$

This is the budget equation for the boundary layer equivalent potential temperature.

$$\frac{\partial h_1}{\partial t} + \frac{1}{r} \frac{\partial \left( r u_1 h_1 \right)}{\partial r} = Q_{b,1} - Q_{1,b} - Q_{1,2}, \tag{4.6}$$

This is the continuity equation of the middle layer.

$$\frac{\partial h_2}{\partial t} + \frac{1}{r} \frac{\partial \left( r u_2 h_2 \right)}{\partial r} = \frac{Q_{b,2}}{\varepsilon} + \frac{Q_{1,2}}{\varepsilon}, \tag{4.7}$$

This is the continuity equation of the upper layer

$$w = -\frac{h_b}{r} \frac{\partial \left( r u_b \right)}{\partial r}, \tag{4.8}$$

The continuity equation of the boundary layer.

$$P_1 = g \left( h_1 - H_1 \right) + \varepsilon g \left( h_2 - H_2 \right), \tag{4.9}$$

This is the hydrostatic equation of the middle layer.

$$P_2 = g\,(h_1 - H_1) + g\,(h_2 - H_2)\,, \tag{4.10}$$

This is the hydrostatic equation of the upper layer.

In the above equations $r$ is the radius, $t$ time, $u$ is the radial wind, $v$ is the tangential wind and $w$ is the vertical velocity at the top of the boundary layer. $H$ is the mean layer depth, $P$ is the kinematic pressure anomaly, $\theta_e$ is the equivalent potential temperature (EPT), $f$ is the Coriolis parameter, $\eta = f + r^{-1}\frac{\partial rv}{\partial r}$ is the absolute vorticity, $g$ the gravitational acceleration and the terms $D_{v,X}$, $D_{h,X}$ and $D_{s,X}$ represent the tendencies of quantity X due to the vertical exchange between layers, the horizontal mixing and the surface fluxes, respectively (see Frisius and Lee (2016) for the equations of these tendencies). The indices b, 1 and 2 denote the boundary layer, middle layer and the upper layer, respectively.

The vertical mass movement in the model are represented by mass fluxes. The upward convective mass fluxes are represented such that they are proportional to the frictionally induced positive vertical velocity at the top of the boundary layer:

$$Q_{1,2} = \frac{1}{2}\left[(\eta - 1) + |\,(\eta - 1)\,|\right] Q_b, \tag{4.11}$$
$$Q_{b,1} = \frac{1}{2}\left[(1 - \eta) + |\,(1 - \eta)\,|\right] Q_b, \tag{4.12}$$
$$Q_{b,2} = Q_b - Q_{b,1}, \tag{4.13}$$

where $Q_b = \frac{1}{2}\left(w + |w|\right)$ and $\eta$ is the entrainment parameter and it is defined as:

$$\eta = 1 + \frac{\theta_{e,b} - \theta^*_{e,2}}{\theta^*_{e,2} - \theta_{e,1}}, \tag{4.14}$$

The asterisk presence of the $^*$ means a quantity that was evaluated when it was saturated with water vapour. $\theta_{e,1}$ is assumed to be constant while $\theta_{e,2}$ is dependent on the depth of the upper layer. $\eta$ gives a measure of convective instability and there is a switch to allow deep convection so when

$\eta$ is greater than 1 we have deep convection. Without deep convection there will be no tropical cyclogenesis. Then, the frictional convergence in the boundary layer is not sufficient to support the cyclone development. As a result $\eta - 1$ gives a measure of deep convection and therefore is the convective instability parameter. To ensure that there is mass conservation in the boundary layer, $Q_b$ yields a downward massflux:

$$Q_{1,b} = -\frac{1}{2}\left(w - |w|\right), \tag{4.15}$$

We would like to test the WISHE paradigm in this model. In O69 convection is brought about by frictional convergence in the boundary layer but Frisius and Lee (2016) have successfully shown through a stability analysis that for certain $\eta$ values there will be no convective activity if there is no friction. This situation means that CAPE can build up in these regions which makes O69 not a suitable model to bring about cyclogenesis under the WISHE paradigm. So to allow for us to incorporate WISHE in our experiments we provide convection that does not have any dependence on frictional convergence thus causing us to modify the mass fluxes for when $\eta > 1$:

$$Q_{1,2} = (\eta - 1)\left(Q_b + \frac{\eta - 1}{\tau_C} h_b\right)\beta, \tag{4.16}$$

$$Q_{b,1} = \left(Q_b + \frac{\eta - 1}{\tau_C} h_b\right)(1 - \beta), \tag{4.17}$$

$$Q_{b,2} = \left(Q_b + \frac{\eta - 1}{\tau_C} h_b\right)\beta, \tag{4.18}$$

$$Q_{1,b} = -\frac{1}{2}\left(w - |w|\right) + \frac{\eta - 1}{\tau_C} h_b, \tag{4.19}$$

where $\beta = \beta(r, t)$ is the precipitation efficiency. It ranges from 0 to 1 where if it is 0 there will be no precipitation and if it is 1 all the condensed water falls out. The $\beta$ parameter was already introduced in the WISHE-type models of Emanuel (1997), Gray and Craig (1998) and Frisius (2006) where it was an important parameter for intensification. It decreased radially where it was greatest in the eyewall and less in the ambient region. Montgomery et al. (2009) has concerns over using a parameter that was only introduced for specific purposes, however we will use this parameter to investigate whether intensification with zero CAPE occurs only due to the presence of this parameter.

Since $\beta$ is considered to be a fluid property of the middle layer it can be expressed as:

$$\frac{\partial \beta}{\partial t} = -u_1 \frac{\partial \beta}{\partial r}, \tag{4.20}$$

where $u_1$ is the radial wind of the middle layer and the condition $\beta = 0$ has been applied at the lateral boundary. The surface transfer coefficients for momentum and enthalpy are equal $C_D = C_H = 0.0005 \left(1 + \frac{V_b}{8.33}\right)$ where $V_b$ is the wind speed in the boundary layer. These calculations are based on a bulk scheme. A similar method was used in the low-order model.

We use a simplified expression for CAPE:

$$E_{C,O} = \frac{C_p}{\alpha} \left(\theta_{e,b} - \theta^*_{e,2}\right), \tag{4.21}$$

where $C_p$ is the specific heat at constant pressure and $\alpha = 10$ a thermodynamic constant which was introduced by Ooyama. For deep convection $E_{C.O}$ needs to be positive.

Since the Ooyama model assumes upright convection we can derive the thermal wind balance equation in physical space:

$$\frac{v_1^2}{r} - \frac{v_2^2}{r} = -\frac{C_p}{\alpha}\frac{\partial \theta_{e,b}}{\partial r} + \frac{\partial E_{C,O}}{\partial r} \tag{4.22}$$

where $v_1$ and $v_2$ represent the middle and upper layer tangential winds, respectively.

## 3. Experiment Set Up

We use the unbalanced dynamics in O69 where the vortex is axisymmetric. The numerical model uses 2400 stretched radial grid points where the distance is 250 metres in the inner part of the domain while in the outer domain part the grid point distance is 2500m. To avoid the build up of sharp gradients we allow horizontal diffusion where the diffusion coefficient is $\nu_h$ = $1000 m^2/s$. The tangential profile of the initial vortex takes the form of

$$v_j = 2V_{max,0}\frac{r/r_0}{1+(r/r_0)^2} \text{ for } j = 1, b \tag{4.23}$$

where $V_{max,0} = 10$m/s is the maximum tangential wind and $r_0 = 50$km is the radius of maximum wind. Initially, there is no flow in the upper layer and only the REF_OOYAMA experiment adopts the O69 frictional convergence. Below describes each experiment:

- REF_OOYAMA: Adopts unbalanced dynamics in both the boundary layer and the free atmosphere. To ensure there is zero CAPE initially $\theta_{e,b}$ = $\theta^*_{e,2}$. The efficiency parameter $\beta$ is set to one we can ignore the effects of a radially decreasing efficiency parameter.

- REF: Is identical to REF_OOYAMA except that the additional convective exchange is triggered besides the convection being triggered by frictional convergence. The timescale for convection $\tau_C$ = 24 hrs.

- HIGHCAPE: Same set up as REF except the initial $\theta_{e,1}$ and $\theta^*_{e,2}$ are decreased by 5 K and 10 K, respectively.

- STABLE: Same set up as REF except the initial $\theta_{e,1}$ and $\theta^*_{e,2}$ are increased by 5 K and 10 K, respectively.

- WISHE: Adopts the additional convective scheme but with $\tau_C$ = 1 min.

- WISHE_DRY: Uses the radially decreasing precipitation efficiency parameter where the profile for $\beta$ at the beginning is:

$$\beta\left(r, t=0\right) = \frac{1}{2} - \frac{1}{2}tanh\left(\frac{r-r_0}{10km}\right). \tag{4.24}$$

  This profile will produce a dry region beyond the initial radius of maximum wind, $r_0$, thus creating the necessary conditions for intensification under the WISHE paradigm. In the WISHE models that uses the efficiency parameter it decreases radially at the radius of maximum winds.

Table 4.1: Parameter values for the Ooyama model experiment REF.

| Notation | Value | Meaning |
|---|---|---|
| $\epsilon$ | 0.9 | Upper to middle layer density ratio |
| $h_b$ | 1 km | Boundary layer height |
| $H_1$ | 5 km | Mean middle layer depth |
| $H_2$ | 5 km | Mean upper layer depth |
| $f$ | 5x$10^{-5}$s$^{-1}$ | Coriolis parameter |
| $\theta_{e,1}$ | 332 K | Middle layer EPT |
| $\theta_{e,2}$ | 342 K | Ambient value of upper layer EPT |
| $\theta^*_{e,s}$ | 372 K | Ambient value of sea surface EPT |

# 4. Results

Figure 4.2 shows the time evolution of the maximum tangential wind in the middle layer (top graph) and the radius of the maximum tangential wind in the middle layer (bottom graph). As per Ooyama (1982) intensification is a result of inflow in the deep-layer, hence the maximum tangential wind will not be in the boundary layer but in the middle layer. As a result our parameter of interest is $v_1$. From the top graph in figure 4.2 REF_OOYAMA and HIGHCAPE both have an early onset of intensification which are almost identical for the first 75hrs. The STABLE experiment shows a very slow and steady development but only ends up attaining tropical storm strength winds at the end of the simulation. The vortex in the WISHE experiment has a slow intensification also, and it too only manages to attain tangential wind of a tropical storm strength. The REF experiment develops to maturity and then after 170hrs it starts to decay. HIGHCAPE shows a similar result. It develops quickly, attains high wind speed and then it decays after 110hrs. As a result there is no steady state for REF and HIGHCAPE. This is not a surprise since O69 stated that any TC simulated with this model will not approach a steady state. The experiment WISHE_DRY decays. There is no intensification. The WISHE_DRY experiment employs a radially decreasing efficiency parameter that causes a dry region to develop beyond the initial radius of maximum wind. This parameter is what causes the vortex to decay because we see that there is some intensification in WISHE, albeit slow and small. REF_OOYAMA has the highest tangential wind which is maintained throughout the simulation until near the end where it starts to decay. Therefore, it can be inferred that convection that arises due to frictional convergence supports the most rapid intensification and thus provides a better maintenance of the vortex.

The radius of maximum wind (RMW) in the middle layer contracts during intensification. We see this in REF, REF_OOYAMA and HIGHCAPE. It is not a surprise that both REF_OOYAMA and HIGHCAPE radii contract in the same way because their tangential wind evolution was almost identical for the first part of more specifically during the development stage. The experiment STABLE has an increase in RMW after 50hrs. This increase lasts until 145 hrs and then the radius begins to contract. The WISHE experiment has an increase in RMW that has a steady growth outward from the centre of the storm. This outward migration with no amplification is consistent with the results of Frisius (2006) where it was revealed that for amplification to occur there needs to be a negative gradient of the efficiency parameter. Without it there will be no intensification under the WISHE mechanism accompanied by an outward migration of the RMW. However, WISHE_DRY does employ the radially decreasing precipitation efficiency parameter and yet there was still an outward migration of the RMW with no amplification. This result means that intensification in the Ooyama model cannot be caused by WISHE. Based on the WISHE results we can infer that the Ooyama model needs a significant amount of CAPE to allow intensification. Even if there is little CAPE which was the case in WISHE because the additional convective scheme allowed for convection to take 1 minute to stabilise the atmosphere. During this very short time there was CAPE albeit small. The WISHE mechanism had a small increase but the lack of CAPE made it difficult for the vortex to attain any real tangential wind strength.

The question remains whether the problem is the choice of model. The drawback with WISHE models is that they neglect the entrainment of low entropy air into the boundary layer while Ooyama's model adopts a time invariant middle layer entropy.

Figure 4.3 shows the radial profile of of CAPE ($E_{c,o}$) and the tangential wind ($v_1$) for REF_OOYAMA, REF, HIGHCAPE and STABLE experiments at the time when the intensity reaches 50% of its time maximum. The radial profile of CAPE for all experiments shows that near the centre CAPE is high and then it dips just before the RMW. This however, is not seen in the STABLE experiment. This dip has a very narrow radial range and the source of it being zero has to do with the fact that it is located inside of the eyewall where downdraughts carry low entropy air into the boundary layer cause the CAPE to be lowered in this areas. In the STABLE experiment we see a dip just before the RMW but it is non-zero. The gradient of the tangential wind for the other three experiments is very steep when moving towards the maximum winds but then the tangential wind falls off exponentially. This is characteristic for a tangential wind profile in a tropical cyclone. REF_OOYAMA has the most intense winds at its RMW which is expected since it had the highest maximum tangential wind speed. There is a significantly higher amount of CAPE in the inner region of the RMW of REF_OOYAMA. This may have to do with the fact that all the other experiments switched on the additional convective exchange while REF_OOYAMA used the frictional convergence, thus allowing for CAPE to accumulate in the inner core.

# 5. Summary

We used a simple axisymmetric model with unbalanced dynamics to study whether the WISHE paradigm for intensification is justified in excluding CAPE from its explanation for intensification. To test the WISHE theory in the modified 069 model, we needed to add an additional convective exchange scheme that does not depend on frictional convergence. The original O69 model cannot support the WISHE mechanism since CAPE can still build up in regions where there is no frictional convergence. To ensure there is no build up of CAPE, the original mass fluxes are modified where a timescale for convection is included. This timescale allows one to make a switch from convection triggered by frictional convergence to convection without frictional convergence. Through a series of experiments where we show the time evolution of the maximum wind speed for experiments where convection is triggered by frictional convergence, the additional convective exchange is used, adjustments are made to the initial equivalent potential temperature (EPT) of the middle layer and the saturated EPT of the upper layer to produce a high cape and stable environment. To better understand the impact CAPE has on intensification we looked at radial profiles of it together with the profile of tangential wind at the time when the intensity exceeds 50% of its time maximum.

The results reveal that convection triggered by frictional convergence provides the best conditions

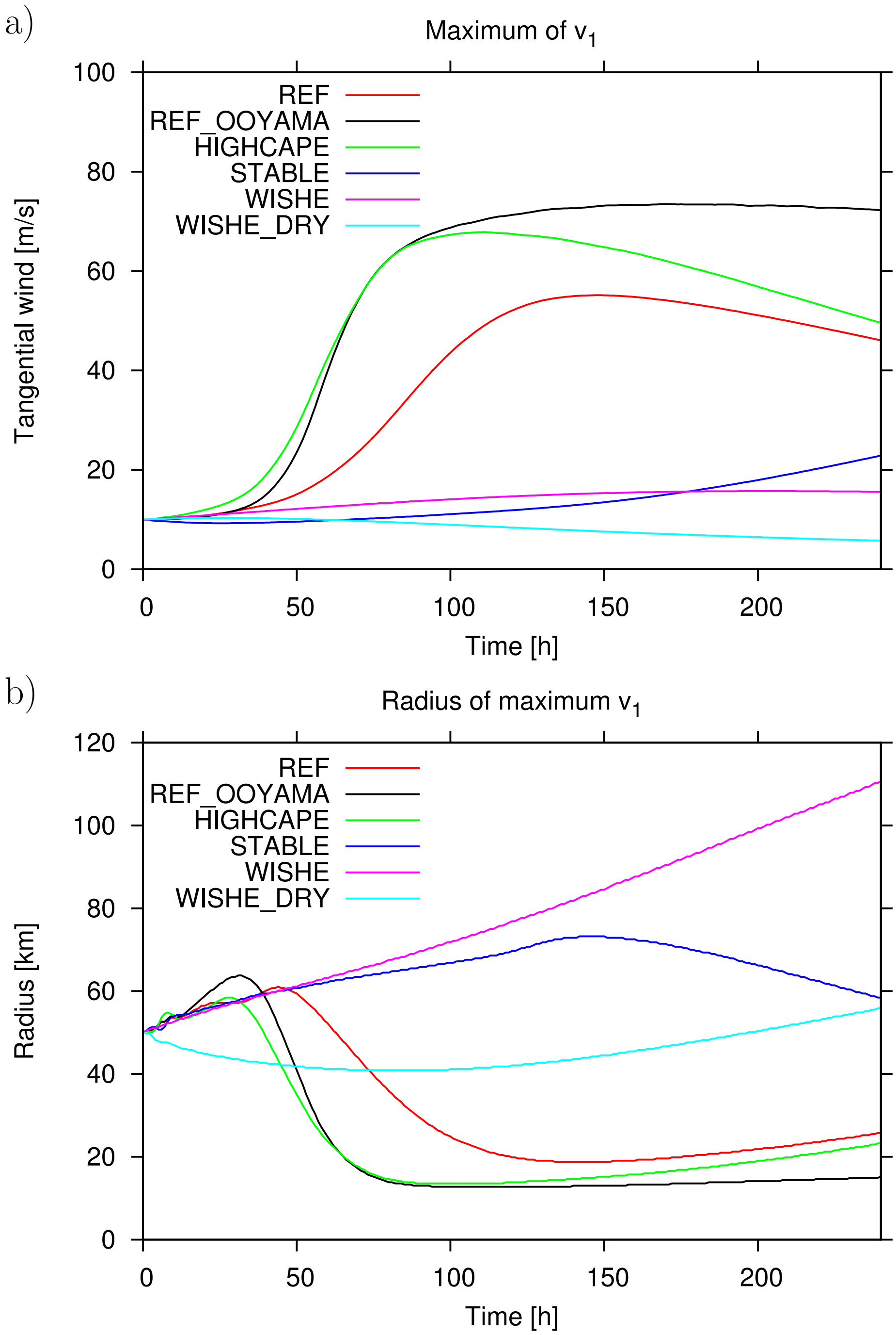



Figure 4.2: a) Maximum middle layer tangential wind as a function of time for the experiments REF (red curve), REF_OOYAMA (black curve), HIGHCAPE (green curve), STABLE (blue curve), WISHE (magenta curve) and WISHE_DRY (light blue curve). b) as in a) but the radius of maximum $v_1$ is shown.

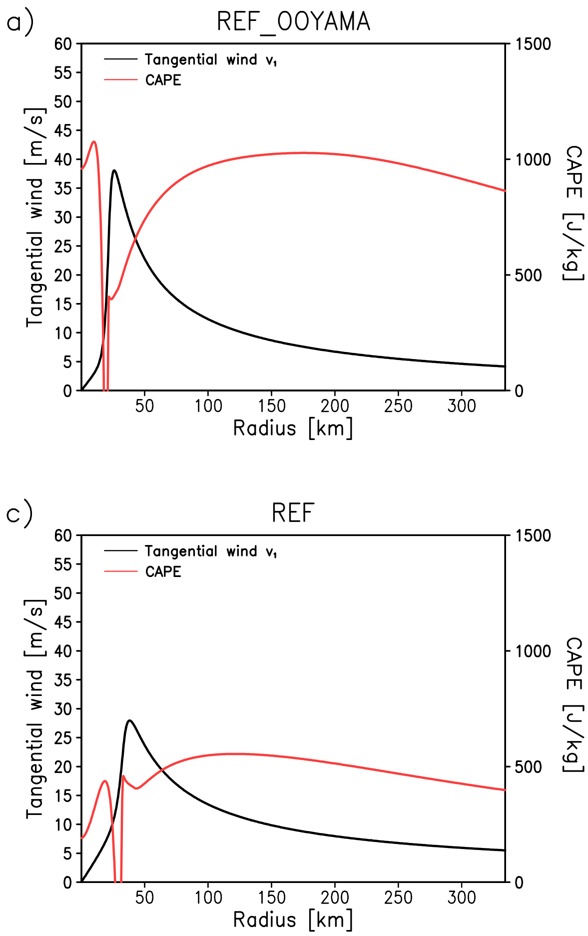



Figure 4.3: Radial profiles of tangential wind $v_1$ (black curve) and CAPE $E_{C,O}$ (red curve) in the Ooyama-model at the time when the intensity reaches 50% of its time maximum for the experiments a) REF_OOYAMA, b) HIGHCAPE, c) REF and d) STABLE.

for the vortex to develop and be maintained. The HIGHCAPE experiment produced a strong vortex in the earlier part of the simulation but then it decayed like the REF. The WISHE and WISHE_dry experiments proved that adjustments need to be made to the WISHE mechanism in order for it to fully explain intensification. The WISHE experiment employed a 1 minute timescale to convection and there was a very slow and steady increase while WISHE_DRY adopted the radially decreasing precipitation to a facilitate its intensification and yet there was no intensification. The vortex decayed instead. The purpose of the radially decreasing efficiency parameter is to address the fact that the WISHE mechanism cannot support intensification unless there are restrictions to the heating which are brought on by the surface fluxes. Frisius (2006) explained this because it was noted that the WISHE mechanism alone cannot support amplification. This further restriction is the radially decreasing efficiency parameter which the WISHE_DRY but yet there was still no amplification. Instead the vortex decayed. As a result of this the question arises if there are limitations in using the Ooyama model to study intensification using the WISHE mechanism. We are aware from the previous section the limitations of WISHE-type models. They neglect entrainment of low entropy air into the boundary layer which is a serious issue. However, Ooyama models have limitations too. In the Ooyama model the assumption is made that the middle layer entropy does not change. This too can have implications. The convective adjustment scheme we used in this model is very similar to that of other Ooyama-type models such as Zehnder (2001) and Schecter (2011) which use a quasi-equilibrium scheme. In this scheme they relate the convective mass fluxes to the disequilibrium between the middle layer and the boundary layer so a decrease in the timescale will bring the atmosphere to a moist stable and not moist neutral state which is what is present in the WISHE model. Zhu et al. (2001) used a minimal three-level tropical cyclone model with different convection schemes to determine if certain features of a developing cyclone are sensitive to a particular parameterization. In their study the cyclone that used the modified Arakawa scheme which relates the convective mass fluxes to CAPE develop much more slowly that the modified Ooyama scheme which relies wholly on convection parametrization through convergence. This too agrees with our result where the REF_OOYAMA experiment that relies on frictional convergence attained the highest intensification.

We further explored how CAPE was radially distributed throughout the cyclone. REF_OOYAMA had the greatest amount of CAPE which has to do with the fact that in this case CAPE can accumulate if there is no convergence. The other experiments adopted the convective adjustment scheme with a time scale of 24hrs. So convective processes were able to reduce the CAPE quantity in those experiments except STABLE which had the least amount of CAPE. The Ooyama model provides a very useful and easy model to study CAPE but there are limitations and these need to be taken into account when evaluating the results.

# Chapter 5

# Cloud Model 1

In this chapter we use a more complex model to better understand the impact CAPE has on the intensification process. We use the three-dimensional non-hydrostatic, non-linear, time dependent numerical model, CM1 version 17 (in Lee and Frisius (2018) version 18 was used but only for selected experiments). This model was developed by George Bryan at NCAR (see Bryan and Fritsch (2002) and Bryan and Rotunno (2009)) for the purposes of research, therefore it can only be used for idealised simulations and not for real cases. Its primary purpose is to study and investigate atmospheric phenomena but on a small scale such as thunderstorms. This model is ideal for studying TC intensification because it resolves for all the necessary moist processes that provide and support tropical cyclone (TC) intensification. The model can simulate and present results in three configurations: 2D, 3D and axisymmetric. In our experiments we use the 3D and axisymmetric configurations to show that differences do occur in these dimensions and as a result they affect the way a cyclone develops. We do not however, investigate nor study the source of these differences. To better understand why TC intensification differs in three dimensions from axisymmetric please read Persing et al. (2013) which gives a full account of the source of the reduced intensification in 3D. They also use CM1 but version 14 to study the intensification in both configurations. Knowing that intensification differs depending on 3D or axisymmetric setup is important because real tropical cyclones as Persing et al. (2013) pointed out are not axisymmetric. Only very intense storms can achieve axial symmetry but this is only present in the inner core (Montgomery and Smith (2014); Persing et al. (2013)). This is cause for concern since some TC simulations and studies treat the cyclone as if it has axisymmetric dynamics. Intensification is not entirely based on axisymmetric dynamics, and as a result there are limitations on how we can interpret the information we get from studies that use axial symmetry. From our intensification theory section, the VHT route provides an explanation for intensification that incorporates asymmetric features (see Montgomery and Smith (2014)), so those researchers are trying to better our understanding of TC from a more realistic point of view. Sang et al. (2008) has highlighted the growing interest the TC community is having with the role asymmetries play in the intensification process. In that study intensification was studied in three dimensions and from their results it revealed that these deep convective vortex structures dominate throughout the intensification process. Unlike Sang et al. (2008) we are using a 3D model to better understand intensification through a series of sensitivity tests. The sensitivity tests will reveal the impact CAPE has on intensification in both an axisymmetric and 3D configuration. The results show that the cyclone is definitely sensitive to the presence of CAPE which indicates that CAPE has a role and an impact on the way the cyclone develops. Using a 3D model to study TC intensification moves us one step closer to understanding the thermodynamics and dynamics of intensification in a real storm. TCs in real life are not axisymmetric. The asymmetries that are superimposed on the axisymmetric vortex impacts convection and intensification. Due to the way CAPE is generated we investigate the effects having CAPE being in the ambient environment or having elevated and lowered values of it during the intensification process.

# 1. Model Configuration

All experiments are performed using CM1 version 17. The governing equations for CM1 have eleven prognostic variables: the three velocity components $(u, v, w)$, the non-dimensionless perturbation Exner function for pressure $(\pi^{'})$, perturbation potential temperature $(\theta^{'})$, the mixing ratios for water vapour $(q_v)$, rain $(q_r)$, cloud ice $(q_i)$, snow $(q_s)$ and graupel $(q_g)$.[1] The Coriolis parameter is set to $5$ x$10^{-5}s^{-1}$ so an f-plane is assumed. The boundary conditions are periodic in all directions. The sea surface temperature is kept constant at 28°C.

[1]See ”The governing equations for CM1” on the web site http://www.mmm.ucar.edu/people/bryan/cm1/ for the full equation set.

The default configured namelist for TC simulation is used except for some changes which we will describe here. The radiation scheme was switched off for all experiments. The double-moment scheme by Morrison et al. (2005) which excludes hail was applied. A Smargorinsky-type scheme that uses different coefficients for vertical and horizontal exchange is used to parametrize the micro-turbulent exchange and dissipative heating. A constant value of $C_H = 0.0012$ for the surface transfer coefficient for heat and moisture is employed while a parameterisation based off Fairall et al. (2003) at low wind speeds and Donelan et al. (2004) at high wind speeds is used for the surface drag coefficient,$C_D$:

$$C_D = \begin{cases} 10^{-3} & V_{\mathrm{10m}} < V_0 \\ 10^{-3}\left[1 + 0.07\left(V_{\mathrm{10m}} - V_0\right)\right] & V_0 \leq V_{\mathrm{10m}} < V_1 \\ 2.4 \times 10^{-3} & V_{\mathrm{10m}} \geq V_1 \end{cases} \tag{5.1}$$

where $V_{10m}$ is the horizontal wind speed at the 10 m height, $V_0 = 5$m/s and $V_1 = 25$ m/s. The configuration for the model as stated above is for both the 3D and axisymmetric experiments. In 3D we use a 2km grid spacing with 600 grid points in the horizontal and 40 model levels in the vertical where these grid points have a spacing of 500m. The Klemp-Wilhelmson time-splitting is adopted for the time integration where we use the vertically implicit scheme. We adopt 7s for the large time step and 4 intermediate short time steps. We define CAPE as:

$$E_C = \int_{Z_{LFC}}^{Z_{LNB}} g\frac{T_{v,p} - T_{v,e}}{T_{v,e}} dz\ , \tag{5.2}$$

where $Z_{LFC}$ is the level of free convection in the boundary layer, $Z_{lnB}$ is the level of neutral buoyancy, $g$ is the gravitational constant, $T_{v,p}$ is the virtual temperature of the moist adiabat and $T_{v,e}$ is the virtual temperature of the environment. An open source FORTRAN code is used to compute CAPE.[2]. To obtain a radial distribution of the pressure minimum we did an azimuthal averaging of it. In 2D we use a 1km grid spacing with 1500 grid points in the horizontal and 40 model levels in the vertical where the grid point spacing is 500m. For our axisymmetric experiments we do a coordinate transformation to potential radius space to calculate SCAPE as opposed to CAPE. The calculation of SCAPE is produced from the same FORTRAN code but there is a coordinate

[2]A copy of the code is provided on the website ftp://texmex.mit.edu/pub/emanuel/TCMAX/.

transformation to potential radius space first.

## 2. Experiment Setup

Dunion (2011) sounding provides the basis for us to adjust the temperature profile so that the pre-depression cyclone can be in an environment with high and low CAPE to investigate how it develops in these environments. The Dunion (2011) sounding has a CAPE value of 1922 J/kg which in itself is already a very high value. This may be due to the fact that he neglects condensate loading and virtual temperature effects. The initial vortex is the same as Rotunno and Emanuel (1987) which is given by their equation (37) resulting in the maximum tangential wind of 12 m/s being at a radius of 100km, the outer radius (tangential wind vanishes) is at 412.5km. There is a decrease in tangential winds with height. The decrease goes up to 20km where the sponge layer is and at this layer the winds are zero.

To adjust the temperature profile we added a horizontally uniform temperature anomaly:

$$T' = \Delta T_t \frac{z}{10000\text{m}} , \tag{5.3}$$

where $z$ is the height and $\Delta T_t$ is the temperature anomaly at a height of 10000m. Note that $T'$ increases or decreases further for larger heights.

The description of each experiment is as follows:

- REF: This is the reference experiment and the model is run with no modification to the sounding.
- COOL10K: The atmosphere is cooled by setting the temperature parameter to $\Delta T_t = -10$ K.
- COOL5K: The atmosphere is cooled by setting the temperature parameter to $\Delta T_t = -5$ K.
- WARM10K: The atmosphere is warmed by setting the temperature parameter to $\Delta T_t = 10$ K.

- WARM5K: The atmosphere is warmed by setting the temperature parameter to $\Delta T_t = 5$ K.

We want to further understand the impact of the CAPE generation rate on intensification. This is achieved by having lower and higher values of the surface coefficient for enthalpy.

The description of each experiment is as follows:

- REF: This is the reference experiment and the model uses the $C_E$ value that is defined above.
- $C_E$ x 4: The transfer coefficient is quadrupled.
- $C_E$ x2: The transfer coefficient is doubled.
- $C_E$ x 0.25: The transfer coefficient is quartered.
- $C_E$ x 0.5: The transfer coefficient is halved.

For the 3D experiments in both cases we append '3D' at the end of the experiment name to distinguish it from the axisymmetric experiments.

# 3. Results

*a. Tangential Wind Experiments*

Figure 5.1a shows the time series of the maximum tangential wind speed ($V_{max}$) for the perturbed ambient temperature simulations in axisymmetric configuration while figure 5.1b shows the same thing for 3D configuration but the wind speed is of the aximuthally averaged tangential velocity. The black circles on the graphs is the time when the intensification was the greatest. To determine this value an eyeball assessment was done since exact calculation is misleading due to the noise in the data. Since the focus of our study is intensification -the development stage - we are only studying the first 6 days of simulation as opposed to the characteristic 10 days. When comparing the experiments in the axisymmetric configuration there is a minor dip in intensification that lasts for about 7 hrs for COOL10K and COOL5k but around 10 hrs for REF and WARM10K and WARM5K. During this time the cyclone is still developing its structures and surface friction dominates at this point until the air becomes saturated and then intensification increases, some more than others. We notice that altering $\Delta T_t$ by 10 K and 5K did not produce different results. Both experiments evolved almost identically. The black circles indicate that for both COOL10K and COOL5K the times when intensification was the greatest are not that far apart. COOL10k has the greatest intensification at T = 36hrs while COOL5K has its greatest intensification at T=43 hrs. It takes less than two days for these experiments to attain maximum intensity but around 85hrs they both stabilise. In the WARM10K and WARM 5K we see that the vortex was very slow in developing. WARM10K maintains tropical storm winds near the end of the development phase with its greatest intensification occurring at T= 115hrs. It never develops hurricane force winds. WARM5K attained its greatest intensification at T=123hrs. Although, WARM5K attained its greatest intensification after WARM10K it still produced stronger winds and managed to attain hurricane force winds nearing the end of its intensification phase just before it stabilises. According to the Saffir-Simpson scale any hurricane wind speeds that exceeds 70m/s (for 1-minute maximum sustained winds)[3] is a category 5 hurricane. We see that REF, COOL10K and COOL5K all maintain wind speeds in that range at maturity.

Looking at figure 5.1b we see a very similar pattern emerge. Both COOL10K and COOL5K evolve almost identically. Even their times of maximum intensification are the same at T= 53hrs. We also notice that the rates of intensification for these two experiments are very similar to that of REF. COOL10K and COOL5K have an earlier onset of intensification. After 10hrs in all experiments the cyclone begins to develop. In COOL10K and COOL5K there is a spike just after the characteristic dip in intensification. The spike in wind speed is much greater in COOL10K. the tangential wind drops significantly where it maintains more level increase. This spike is even greater than in the axisymmetric counterpart. Once again we see that both WARM10K and WARM5K develop very slowly where WARM10K only becomes a tropical depression. The time of greatest intensification is for WARM10K and WARM5K at T= 118 hrs and 105 hrs, respectively. Comparing these results to that in Lee and Frisius (2018) who conducted the exact same experiments on the same platform and the same model but a different version of it,our results are similar with only minor differences. In their experiments their LOWCAPE corresponds to our WARM5K and their HIGHCAPE corresponds to our COOL5K. They provided initial vertical temperature profiles for their CM1 experiments along with the moist adiabat that was a result of the initial condition of

[3]see https://www.nhc.noaa.gov/aboutsshws.php for more information

their REF experiment.

When comparing figure 5.1a to 5.1b we notice that the intensification begins earlier and is stronger in the axisymmetric configuration. In the WARM10K and COOL5K experiments the most rapid rate of increase happens from 27hrs to 90hrs in the axisymmetric experiments but in 3D configuration it ranges from 30hrs to 70hrs and there is a steady and more gentle increase. However, nearing the end of the phase and entering maturity REF, COOL10K and COOL5K all maintain the same intensity level. This is more noticeable in the 3D configuration and less so in axisymmetric but their final intensities in these simulations are very close. These final intensities being in close agreement supports the claims made by Persing and Montgomery (2005) where they stated that environmental CAPE has no influence on the final intensity of a tropical cyclone and as such any maximum potential intensity (MPI) theory- a theoretical upper limit in intensity that a hurricane can attain - should not consider CAPE. In that study they conducted a series of sensitivity tests using the axisymmetric cloud resolving hurricane model of Rotunno and Emanuel (1987). They used the Jordan (1958) sounding that has 1898J/kg of CAPE. This sounding was cooled by up to 2K in the mid levels to provide the High CAPE run and the set up for the low CAPE run was as per Rotunno and Emanuel (1987) where they run the model without any vortex and no radiational cooling and then use the temperature and moisture profiles that emerge after 12hrs. This creates a neutral state. In all three cases, the simulated cyclone's steady state was largely insensitive to CAPE. Our WARM10K and WARM5K are our low CAPE experiments and they do not attain the same final intensities as those of REF and COOL10K and COOL5K. So this contradicts Persing and Montgomery (2005) but this may be due to the method in which they set up their low CAPE simulation. In their low CAPE experiment adjustments are made to both the temperature and moisture profiles while in our two experiments that have low CAPE we only adjusted the temperature sounding . As a result our relative humidity values in the mid-troposphere is smaller than that of REF, COOL10K and COOL5K. Another source of contradiction could be that the model physics in the model of Rotunno and Emanuel (1987) is very different from that of CM1.

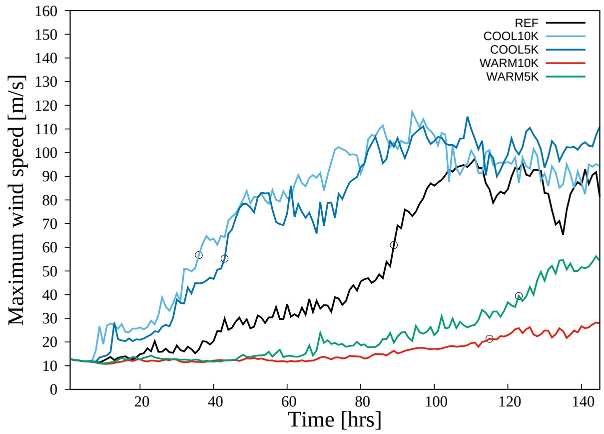
We further investigate CAPE's contribution to intensification by adjusting the surface transfer coefficient of enthalpy fluxes in the model. Emanuel (1986) has stated the source of energy for TCs is the moist enthalpy flux of the sea. Ooyama (1969) has proven that TCs need the fluxes of moist enthalpy to develop and for maintenance. Since this is such a critical parameter in intensification and due to evaporation it generates CAPE, we further investigate to see how adjustments to the flux affect the amount of CAPE which in turn affects intensification. In figure 5.1c where the results of these experiments in the axisymmetric model are shown, we see that very strong and intense vortices develop for when the value of $C_E$ is quadrupled and doubled. Intensification is very intense in $C_E$ x 2 and more so in $C_E$ x 4. From 8hrs to 30hrs the intensification is a little steep and intense in $C_E$ x 4. Then after 30hrs, the curve becomes very steep indicating that the intensification has become very intense and strong. This lasts until T=50hrs when it slows down and stabilise after 105 hrs. $C_E$ x 4 attains its greatest intensification at T=38hrs while $C_E$ x 2 attains it at T= 50 hrs. $C_E$ x 2 follows a similar profile as $C_E$ x 4 but its intensity is less but not by much. In $C_E$ x 4 the onset of the steep and strong intensification is much faster by 10hrs. Both $C_E$ x 0.25 and $C_E$ x 0.5 took longer to develop. Their onset for intensification was slow. They both started around 60 hrs and after 100 hrs $C_E$ x 0.5 had a quick and steady increase in winds where it became a full blown hurricane while the increase in tangential winds was slower and more steady in $C_E$ x 0.25 although hurricane force winds appear at the end of the simulation. $C_E$ x 0.25 attained its greatest intensification at T= 112 hrs while $C_E$ x 0.5 attained it at 100hrs. This is a much slower and longer time than in $C_E$ x 2 and $C_E$ x 4. The REF, $C_E$ x 2 and $C_E$ x 0.5 experiments end their runs with wind speeds that are not too different while $C_E$ x 4 has a little bit higher wind speed and that of $C_E$ x 0.25 is very much less and different.

Figure 5.1d we look at how the vortices with the modified fluxes evolve through time in the 3D setting. The first 40 hrs of REF, $C_E$ x 2, $C_E$ x 0.5 and $C_E$ x 0.25 all reveal the same very low intensity. Only $C_E$ x 4 has an onset of intensification much earlier than the rest. $C_E$ x 4 attains its maximum intensification at T= 42hrs while $C_E$ x 2 attain its at T=54 hrs. Although $C_E$ x 2 had a very slow start it maintained the same rate as $C_E$ x 4. The slower developing vortices of $C_E$ x 0.25 and $C_E$ x 0.5 attain maximum intensification at T= 110hrs and T= 99 hrs, respectively. All experiments produce intense tropical cyclones at the end of the simulation. $C_E$ x 4 and $C_E$ x 2 both at wind speeds that are very close to each other. REF also attains a final wind speed that is close to those two experiments. It is very noteworthy that although the cyclone took very long to develop particularly in the case of $C_E$ x 0.5 and $C_E$ x 0.25, the final intensities were of hurricane strength and if we were to run the model longer, it is quite possible $C_E$ x 0.5 and $C_E$ x 0.25 may have ended closer to the other experiments.

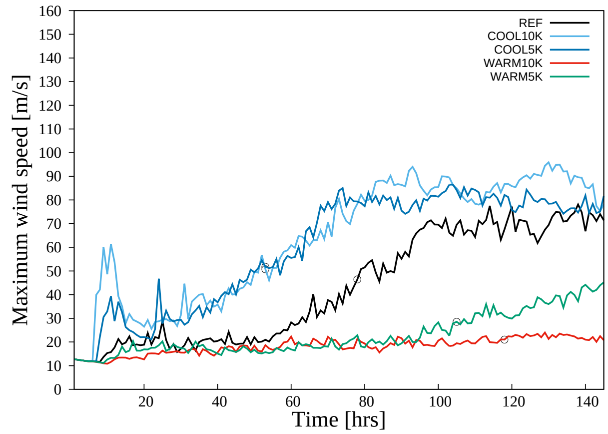
In both axisymmetric and 3D configuration for this experiment the vortices all maintain hurricane force winds near the end of the simulation. Although, the axisymmetric experiments are more intense, the less intense 3D experiments are still able to support a hurricane strength vortex when $C_E$ is modified. Since surface fluxes of moisture and heat are important in the survival of the vortex it is not a surprise that increasing the surface transfer coefficient for enthalpy by doubling and quadrupling produces faster intensifying cyclones. It also no surprise that a reduction in them will result in a reduction of the intensification. When the fluxes are reduced it means less heat and moisture will be entering the air of the developing cyclone. Both Ooyama (1969) and Emanuel (1986) have stressed the importance of this parameter in development and maintenance. The reduction in the fluxes meant less heat and moisture was being fed to the developing cyclone. There was enough moisture to allow a slow development even in the presence of frictional forces which dominate at the very beginning. Reducing this parameter meant that the air just about the cyclone would take longer to become moist and thus there will be a delay in the increase in the tangential winds. For intensification to occur there needs to be saturation of the air. In all these experiments the axisymmetric configuration always had more intense vortices with higher values for their wind speeds. This difference is expected. Persing et al. (2013) has highlighted some of the sources of the differences in cyclone intensification in axisymmetric and 3D configuration. They bring attention to the fact that convection is not organised in concentric circles in the 3D model. As a result there is a reduction in the azimuthally averaged heating rate and radial gradient. Due to the lack of organization of convection, the vortex takes longer to develop which leads to a weaker mature vortex. Our result is consistent with others who studied the differences in axisymmetric and 3D setup such as Persing et al. (2013).

### *b. Profiles of CAPE and SCAPE with Tangential Wind*

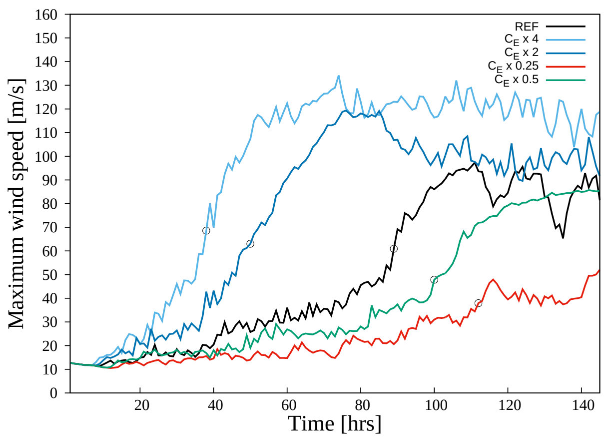
Figure 5.2 shows profiles of tangential wind, SCAPE and time average of SCAPE (where the size of the window of the SCAPE being averaged was 25 data points) for the axisymmetric experiments as a function of potential radius in the left panel at the time of maximum intensification



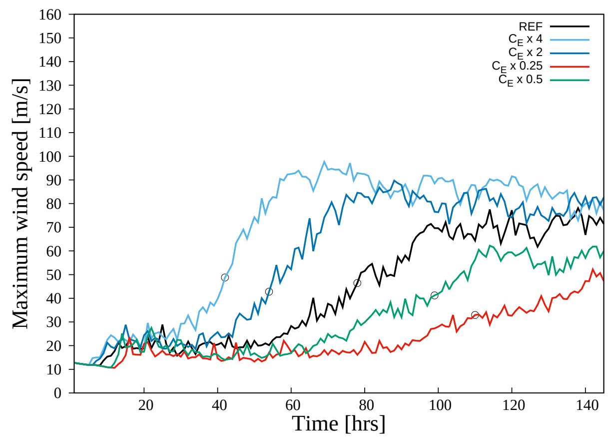

(a) Perturbed temperature experiments for axisymmetric setup.




(b) Perturbed temperature experiments for 3D setup.




(c) Surface transfer experiments for axisymmetric setup.




(d) Surface transfer experiments for 3D setup.

Figure 5.1: Time evolution of maximum wind speed for experiments in the axisymmetric and 3D setups.

and the right panel shows the azimuthally averaged CAPE with wind speed at the time of maximum intensification as a function of radius for the 3D experiments. Due to technical limitations, we did not calculate SCAPE in the 3D experiments. We expect that CAPE and SCAPE to look qualitatively similar. Additionally, SCAPE will always be larger than CAPE due to the fact it lies along lines of slanted angular momentum surfaces, so the presence of CAPE means that SCAPE is also non-zero. Time averaging of SCAPE was done on the axisymmetric results to show a better profile of SCAPE since it was a bit noisy. Also, the axisymmetric results are in Potential Radius (PR) space unlike 3D so that the profiles shapes differ significantly. All calculations for the axisymmetric were done in the new coordinate transformation. This, however does not impede the ability to compare the two. In the axisymmetric results we see that the tangential wind values for REF, COOL10K, COOL5K, WARM 5K are high while the tangential wind of WARM10K is very small. We see that three of the experiments have peaks that attain a significant wind speed value. The REF experiment peaks at 250km with a tangential wind speed of 85m/s. There is a small amount of SCAPE near the centre but as one moves outward, SCAPE drops to zero at around the same time the tangential wind speed is at its greatest. SCAPE in the outward region attains a final value of 2600J/kg. In COOL10K near the centre of the inner core SCAPE is non-zero and it attains a value as high as 2000J/kg. As the wind profile increases SCAPE decreases until 270km where it maintains a minimum for over a short distance and then it increases again. The tangential wind of experiment COOL10K peaks in the presence of high CAPE.

It is interesting to note that although WARM5K maximum wind speed was significantly lower than COOL10K, COOL5K and REF, its tangential wind profile was in the same range as the others. The profile for the time-average of SCAPE shows that as one moves away from the vortex centre we see that SCAPE increases where its maximum is just outside the radius of where the tangential wind profile is at its maximum wind speed. This is true for all except COOL10K and WARM10K. In COOL10K although the SCAPE is not very high there are still significant values within the radius where the wind speed is a maximum. Even at the point where the wind is a maximum there is still some SCAPE. In the case of WARM10K there is no SCAPE present. This is expected since in this experiment the temperature sounding was increased to the point it probably created a moist neutral atmosphere.

The 3D results show similar patterns. The REF_3D experiment (all the experiments in this figure have the 3D appended to their name here) has a sharp increase in the wind speed profile. The gradient is steeper than in figure 5.2a because the physical radius coordinate is used instead of potential radius. The maximum tangential is smaller than in the corresponding axisymmetric experiment. This is also seen in COOL10K_3D and COOL5K_3D. After 30km the wind speed exponential decreases in REF_3D and COOL10K_3D. This is not surprising since both experiments have very similar intensification rates in figure 5.1b. In COOL10K_3D the decrease is a little steeper and faster from 30km to 120km. CAPE is very low within the 60km radius in REF_3D which corresponds to its wind peak. CAPE is very low within the 120km radius of COOL10K_3D. In COOL5K_3D the wind speed peaks at 20km and then has a gentle exponential decline like the other two experiments mentioned above. This experiment has an almost idential intensification rate as COOL10K_3D so it is not a surprise the profile is very similar to COOL10K_3D and REF_3D. However, its CAPE values outside the most intense winds are much higher. This is not surprising since there was actually more CAPE present in the atmosphere due to the fact that the sounding was cooled to a much lower temperature. In WARM10K_3D there is a very small peak and no CAPE present while in WARM5K_3D there was also a very small peak albeit larger than the peak in WARM10K_3D, and there was some CAPE.

Figure 5.3 is the same as figure 5.2 except that these experiments employ different values of the surface exchange coefficient for enthalpy, $C_E$. The axisymmetric results in the left column are also in potential radius space. The $C_E$ x 4 has a very strong maximum at 330km with the tangential wind profile attaining a value of 110m/s. Similar to REF SCAPE increases from zero at the centre maintain values in its time average curve of less than 1000J/kg. As we move further outward SCAPE drops to almost zero near the peak and then it increases rapidly until it attains values close to 3000J/kg. $C_E$ x 2 has the same shape but its values for SCAPE are much less. Near the centre the SCAPE is around 500 J/kg and drops to zero just before the tangential profile reaches its peak and then it rises as one moves further away from the maximum wind speed. The SCAPE in the outward region is just under 3000J/kg. The tangential wind profile peak is at 100m/s. Both $C_E$ x 0.25 and $C_E$ x 0.5 have smaller but still significant tangential winds. $C_E$ x 0.25 attains a maximum tangential wind speed of 43 m/s while $C_E$ x 0.5 attains a wind speed of 60m/s. They both have very low amounts of SCAPE up until 180km from the centre. The SCAPE in both cases radially increases slightly, dips and then increases again up to 1400J/kg in the $C_E$ x 0.25 experiment and 1700J/kg in the $C_E$ x 0.5 experiment.

The wind speed in the 3D experiments is not very intense. All the graphs have the same shape for the tangential wind which is consistent with the time evolution of the maximum wind speed in figure 5.1d. Since they all had similar intensification rates. The tangential wind is weak in $C_E$ x 4 attaining a value just under 40m/s. The CAPE profile dips at the point when the tangential wind is at its highest value at 30km from the centre. Experiment $C_E$ x 2 has a similar wind profile to $C_E$ x 4 but its CAPE profile has a major dip in the region of the cyclone that is far from the inner-core. The CAPE drops to an ending value of 1300J/kg. In both $C_E$ x 0.5 and $C_E$ x 0.25 the wind speed profile is similar to the other experiments. Their CAPE profiles show 2 maxima just like in REF and $C_E$ x 2. They both have ending values for CAPE at 1000J/kg. Except in $C_E$ x 0.5 it maintains 1400J/kg CAPE from 380km until 460km and then drops to 1000J/kg thereafter.

When comparing the results from the axisymmetric with those of the 3D experiments we see that they all have the same pattern where CAPE and SCAPE are minimal within the radius (potential radius) where the wind speed profile is at its greatest. This is expected. The convective processes during intensification especially when intensification is at its greatest will consume the maximum amount of CAPE/SCAPE which is why the areas where the wind speed is at its greatest the CAPE/SCAPE is at its lowest. Having high CAPE/SCAPE outside of the eyewall region (region where the wind speed is most intense) helps to enhance the radial temperature gradient that helps to drive the TC circulation. Having high values of CAPE outside the radius of maximum winds does have a direct impact on intensification. Xu and Wang (2010) investigated the sensitivity of the TC inner-core size to the radial distribution of surface entropy flux and it was revealed that the presence of surface entropy fluxes outside the inner-core supported large CAPE and convection. This results in the development of spiral rainbands. The diabatic heating that is present in these spiral rainbands affects the boundary layer inflow outside the eyewall. The presence of CAPE is related to these rainbands which affect the tangential wind increase.

*c. Top view Profiles*

Figure 5.4 shows snapshots of CAPE and surface pressure for 5 of the 3D experiments at the time when the intensification was the highest (see black circles in fig.5.1b and fig.5.1d). The REF, COOL10K and $C_E$ x4 all have well developed cores where the surface pressure is very low. The isolines are so close you cannot see any white spaces indicating no CAPE. As one moves outwards from the inner core the CAPE increases with increasing radius.In the WARM10K experiment we see that there is virtually no CAPE which correlates well with the other results for this experiment. In REF and in $C_E$x0.25 there is a white ring that separates the inner core form the outer core. The white ring is more pronounced in $C_E$x0.25

# 4. Summary

In this chapter we use a non-hydrostatic cloud resolving model to test the sensitivity of a developing tropical cyclone to different values of environmental CAPE and surface transfer coefficient for enthalpy. To produce these experiments we added a horizontally uniform temperature anomaly where the sounding was cooled and also warmed by 10K and 5K. The preconfigered namelists for both axisymmetric and 3D were used where minor changes were made. We are interested in investigating both, and comparing how they differ but unlike Persing et al. (2013) we do not investigate to understand what causes the differences. Conducting similar experiments but for the surface transfer coefficient for enthalpy provides a means to understand the effect CAPE has on a vortex during intensification. Rotunno and Emanuel (1987) have stated that initial CAPE is not necessary but CAPE can play a vital role in the intensification phase.

Our results show that during the intensification phase we see that the presence of CAPE does affect the intensification of a TC. CAPE enhances the intensification and if there are too low values of it present the cyclone may not develop. We also notice that intensification enhances with increasing surface transfer coefficient for enthalpy because evaporation at the surface adds CAPE to the inflowing air. This CAPE also has an impact on the developing cyclone. In all surface transfer coefficient experiments the model produced TCs that attained hurricane strength winds. When we investigated the radial profiles of CAPE we see that CAPE tends to be lower in the inner core while in the outer core it is very high. The high CAPE values are usually outside of the radius of maximum wind.

When we compare the 3D to the axisymmetric it is very obvious that the cyclone is significantly more intense in the axisymmetric experiments. Even the radial distribution of CAPE is less pronounced in the 3D cases. In the axisymmetric cases we do a coordinate transformation where we are moving along lines of constant angular momentum. Due to the fact that these line tend to flare out to become slantwise CAPE, SCAPE tends to be larger than CAPE. The distribution of SCAPE was significantly different because it is in PR -space.

We saw the distribution of CAPE in a cyclone and its relationship with pressure in the snapshots of CAPE for the 5 selected experiments for the 3D case. As one moves outwards CAPE increases. Towards the centre of the cyclone CAPE is almost nonexistent and this is because the intense convective processes consume much of the CAPE. The pressure isolines are very close in the inner part of the core.

This model provided the right framework to see how the cyclone in axisymmetric and 3D set up is affected by the initial generation of CAPE. We do believe that CAPE plays a very important role in the development of a cyclone.

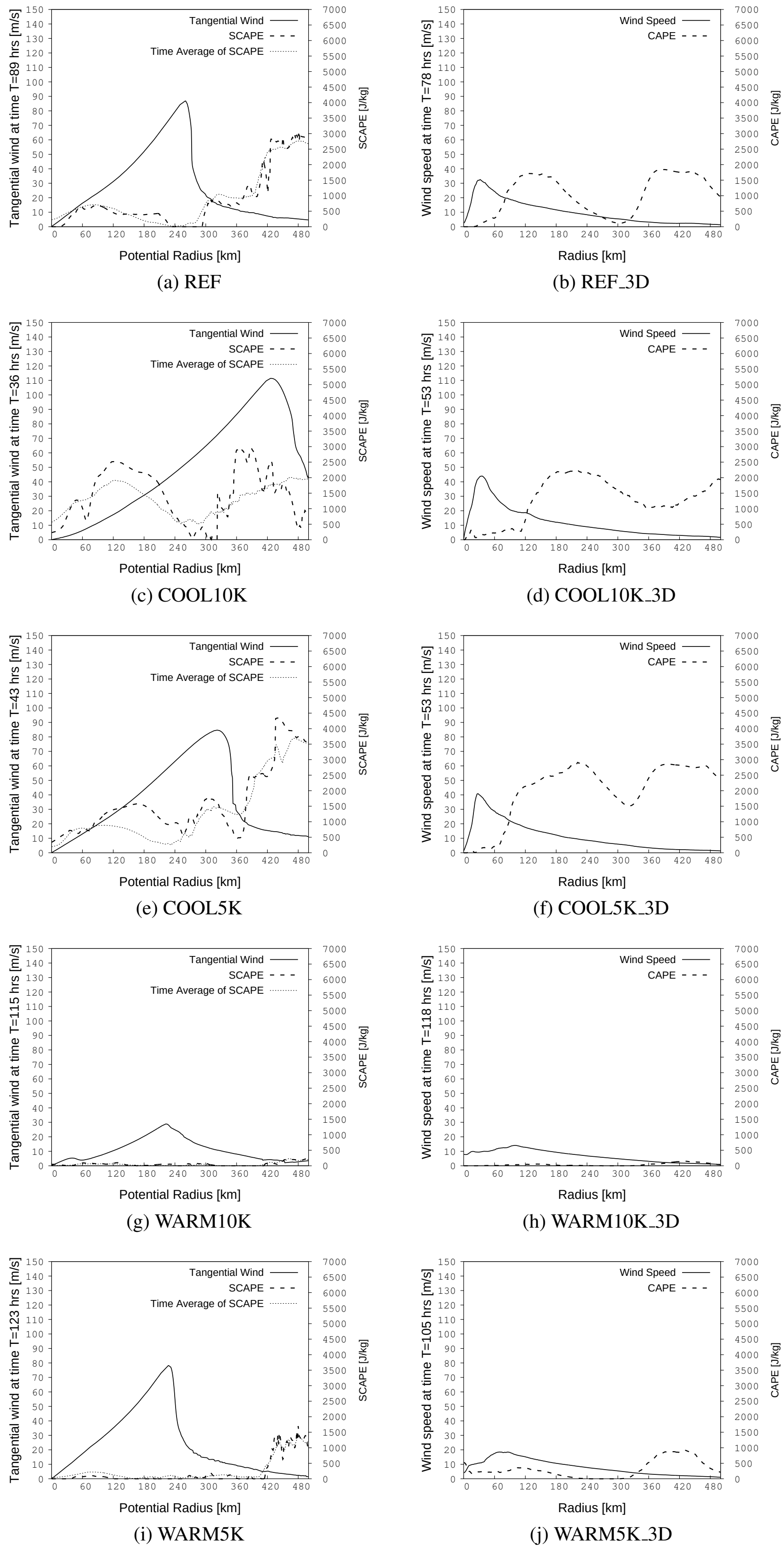
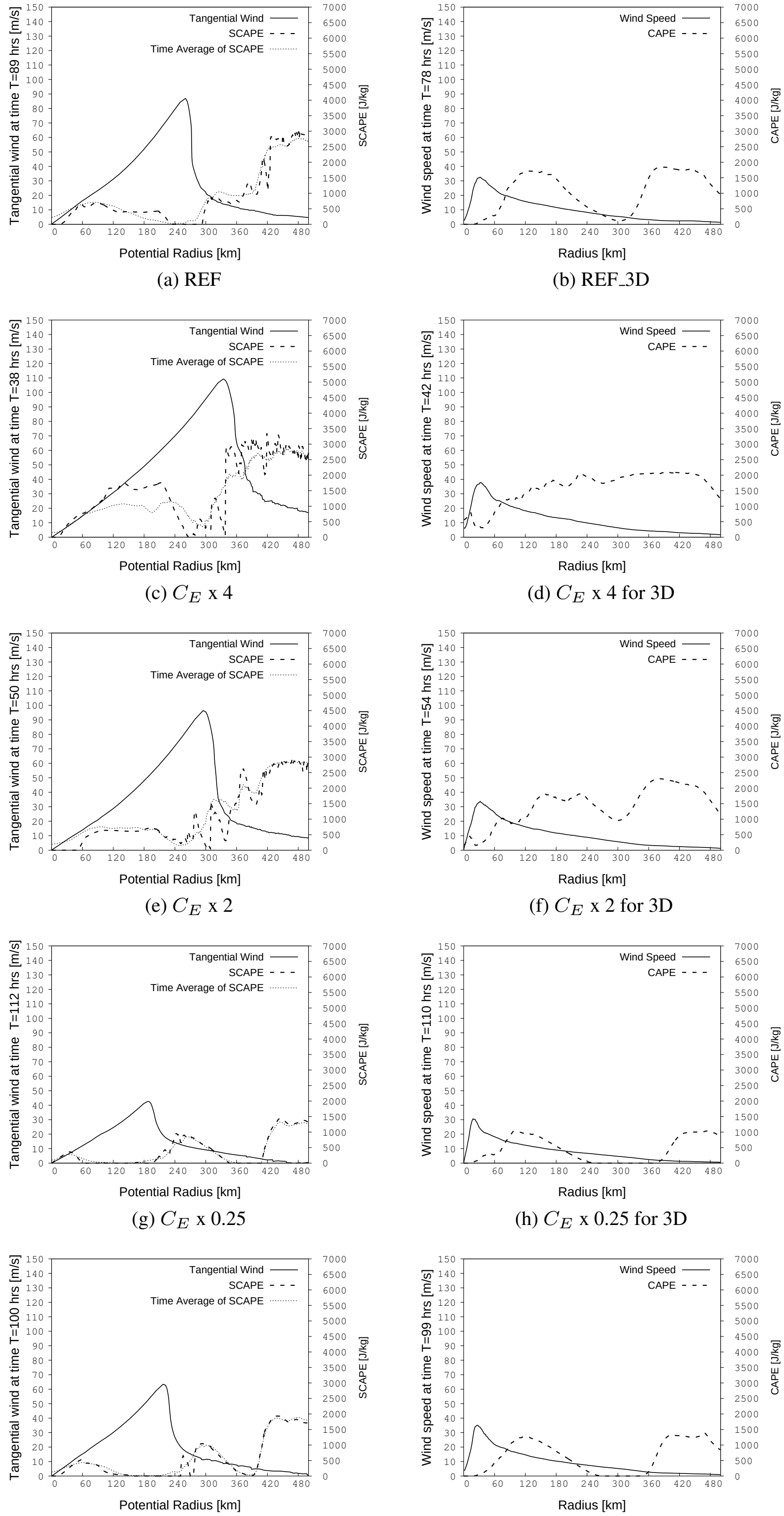



Figure 5.2: Radial profiles of SCAPE and tangential wind in potential radius space for axisymmetric experiments on the left, and radial profiles of CAPE and tangential wind in physical space for 3D experiments at the selected times for the perturbed temperature experiments.




(a) REF (b) REF_3D

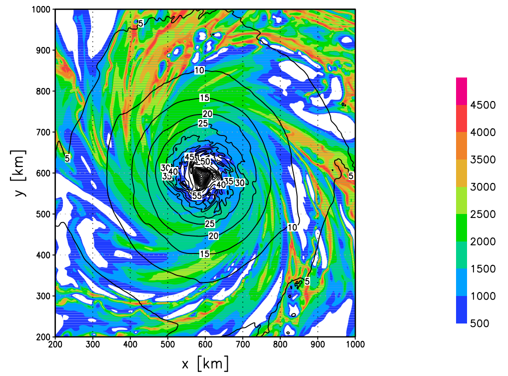
(c) $C_E$ x 4 (d) $C_E$ x 4 for 3D

(e) $C_E$ x 2 (f) $C_E$ x 2 for 3D

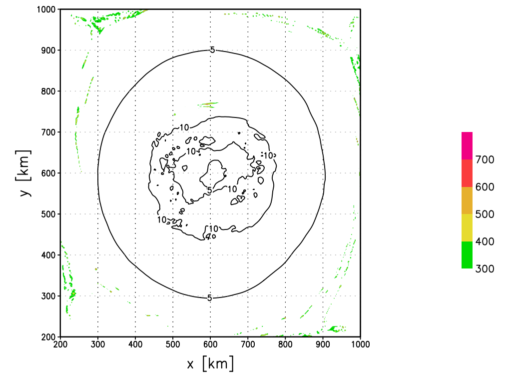
(g) $C_E$ x 0.25 (h) $C_E$ x 0.25 for 3D

(i) $C_E$ x 0.5 (j) $C_E$ x 0.5 for 3D

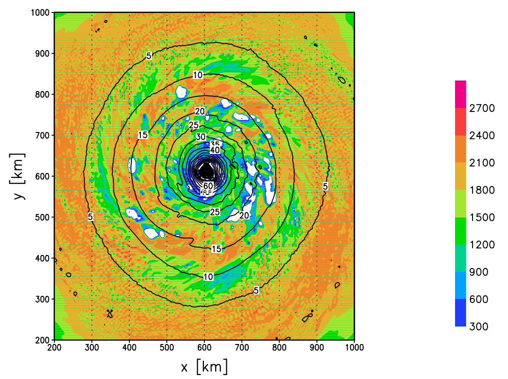
Figure 5.3: Radial profiles of SCAPE and tangential wind in potential radius space for axisymmetric experiments on the left, and radial profiles of CAPE and tangential wind in physical space for 3D experiments at the selected times for the surface transfer coefficient experiments.



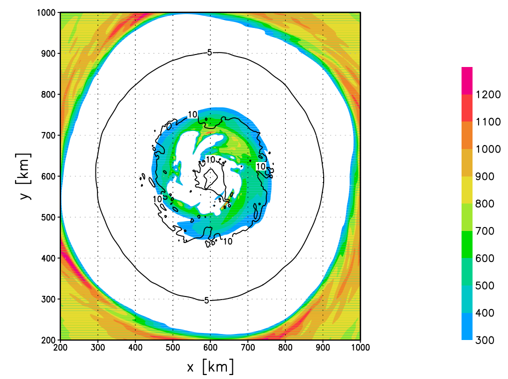

(a) COOL10K



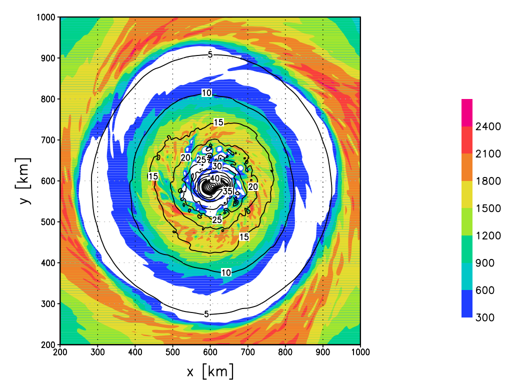

(b) WARM10K




(c) $C_E$ x 4




(d) $C_E$ x 0.25




(e) REF

Figure 5.4: CAPE (coloured shadings) and surface pressure (isolines) for the different 3D experiments at selected times.

# Chapter 6

# Summary and Conclusion

In this study we investigated the role of CAPE in tropical cyclone intensification by using a hierarchy of models. We provided a possible explanation for how CAPE enhances the intensification by relating the presence of CAPE to a build up of high entropy in the boundary layer where it is imported to the eyewall via the secondary circulation. The copious amount of latent heat that is released due to this process increases the negative radial temperature gradient which in turn affects the inflowing air which also affects the gradient wind. We also challenge the validity of WISHE where the mechanism explains intensification with no initial CAPE. In the two simpler models, the WISHE mechanism has always failed to produce a hurricane strength storm even when the radially decreasing precipitation parameter was added. This parameter puts further restrictions on the surface-induced latent heating in WISHE thus allowing vortex amplification, but if the parameter is absent there would be vortex migration with no amplification (Frisius 2006). The fact that the experiment that employed this parameter failed sheds some light on whether the Ooyama model can handle intensification under the WISHE paradigm.

There are two ways convection can be triggered in a tropical cyclone. Firstly, convection can arise out of frictional convergence. The air that converges forces air to rise thus triggering convection. Convection arises also due to processes like precipitation induced downdraughts. From our experiments we see that the forced convection, frictional convergence is the only one that supports intensification. The natural convection does not. In natural convection the convective processes tend to lower the boundary layer air entropy and this low entropy air will eventually be entrained into the eyewall. The convective process will import low entropy air into the boundary layer. The most important finding about our result is the fact that the choice of how setting an extremely small value for the timescale for convection results in no intensification in any experiment. Frisius and Hasselbeck (2009) found agreeing results. They used an axisymmetric tropical cyclone model and found that when evaporation, sublimation and melting are ignored the vortex intensifies very quickly. Additionally, they also had large amounts of CAPE outside of their eyewall. If, however, one includes all the latent cooling processes, intensification would be much slower. Vortex intensification is highly sensitive to CAPE. In the absence of CAPE low entropy air is transported to the eyewall region. WISHE mechanism is incomplete in explaining intensification and through the use of the two simple models, especially in Ooyama not every model has the best set up to be investigating different aspects of intensification. Our sensitivity test further prove that initial CAPE does affect the development of the vortex.

The results from CM1 were also very enlightening. They revealed the impact initial CAPE has on a vortex and how varying amounts of CAPE through adjustments to the enthalpy fluxes also affects tropical cyclones. The intensification rate is related to the radial gradient of CAPE outside the RMGW. The radial CAPE gradient enhances the gradient wind at the surface which leads to higher boundary layer inflow and increased mass fluxes in the eyewall. The experiments for the surface transfer coefficient enthalpy fluxes also revealed that this parameter is also highly related to intensification.

We have been able to successfully show the role CAPE plays in intensification and cast doubt on WISHE as being the answer to intensification while providing an explanation for how CAPE is very instrumental in the development of a vortex. We also bring to one's attention the way convection is implemented can affect your result. In conclusion a radial gradient in CAPE supports TC intensification and should be an important factor when studying intensification.

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# Declaration of Academic Integrity

”Hiermit versichere ich an Eides statt, dass ich die vorliegende Dissertation mit dem Titel: ”The Role of Convective Available Potential Energy in Tropical Cyclone Intensification“ selbstständig verfasst und keine anderen als die angegebenen Hilfsmittel – insbesondere keine im Quellenverzeichnis nicht benannten Internet-Quellen – benutzt habe. Alle Stellen, die wörtlich oder sinngemäß aus Veröffentlichungen entnommen wurden, sind als solche kenntlich gemacht. Ich versichere weiterhin, dass ich die Dissertation oder Teile davon vorher weder im In- noch im Ausland in einem anderen Prüfungsverfahren eingereicht habe und die eingereichte schriftliche Fassung der auf dem elektronischen Speichermedium entspricht.“

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