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Atmospheric and Oceanic Sciences

2015

Tropical Cyclone Activity in an Aquaplanet General Circulation Model

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Princeton, NJ : Princeton University

<http://arks.princeton.edu/ark:/88435/dsp018623j105j>

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TROPICAL CYCLONE ACTIVITY IN AN AQUAPLANET GENERAL CIRCULATION MODEL

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A DISSERTATION

PRESENTED TO THE FACULTY
OF PRINCETON UNIVERSITY
IN CANDIDACY FOR THE DEGREE
OF DOCTOR OF PHILOSOPHY

RECOMMENDED FOR ACCEPTANCE
BY THE PROGRAM IN
ATMOSPHERIC AND OCEANIC SCIENCES
ADVISER: ISAAC M. HELD

JUNE 2015

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Abstract

Understanding the influence of large-scale environmental factors on tropical cyclone (TC) formation is a task of great scientific and societal importance. While much has been learned about the genesis of tropical cyclones over recent decades, a dynamical theory linking the annual global frequency of TCs to the large-scale environment remains elusive. This investigation is motivated by previous studies that have shown that future projections of TC activity will depend critically on the reliability of projections of the change in the pattern of sea surface temperature (SST).

Atmospheric general circulation models are now routinely run at sufficient resolution to enable the internal generation of TC-like disturbances within the simulations and can thus provide an important tool for investigating the behavior of TCs in different climates. This study employs an aquaplanet configuration of the High-Resolution Atmospheric Model (HiRAM) developed at NOAA's Geophysical Fluid Dynamics Laboratory. Through systematically modifying HIRAM's lower boundary condition, this research has explored the impact of changes in the meridional and zonal pattern of SST on the simulated TC activity.

Within these simulations we find that TCs prefer to form near 15°N , along the poleward flank of the Intertropical Convergence Zone (ITCZ). The underlying meridional SST pattern influences the position of the ITCZ, which in turn influences the frequency of TC genesis: the number of TCs increase as the mean ITCZ latitude shifts poleward. The changes in frequency can be related to both the upward mass flux and ambient vorticity of the large-scale environment.

The evolution of the TC along-track intensity was also explored. The intensity of TCs decrease as the separation in the mean latitude of the maximum SST and ITCZ decreases. The average lifetime maximum intensity of TCs is positively related to the period of intensification and the latitude at which the maximum intensity is acquired.

With the introduction of a zonally asymmetric SST perturbation the TCs preferentially form over the regions of relatively warm SST. As the mean zonal gradient of SST increases the TC occurrence density increases over the warmest regions, however the total global frequency remains relatively insensitive to the zonal SST gradient.

Acknowledgements

It is difficult, indeed perhaps impossible, to adequately express my gratitude to the many people who have supported me, both professionally and personally, during this period of study and research as a graduate student at Princeton University. These initial remarks serve only to scratch the surface, but I hope in the fullness of time to be able to personally convey my deep appreciation to each and every one of you.

I have thoroughly enjoyed being a part of the Program in Atmospheric and Oceanic Sciences, and the tremendous opportunity that I have been afforded to learn from the members of the faculty cannot be overstated. In particular I want to thank Dr. Geoff Vallis, Dr. Gabriel Lau, Dr. Stephan Fueglistaler, Dr. Steve Garner and Dr. Gabriel Vecchi for their help in guiding my research at various points along this journey. A huge debt of thanks must also go to the administrative staff, especially Anna Valerio and Laura Rossi, who continue to ensure the students and faculty in the program have all the support we need.

Most importantly I wish to thank my adviser, Dr. Isaac Held. Being able to study under such a highly respected scientist and leader in our field has been the utmost privilege and a rare opportunity that I will cherish for years to come. Aside from his academic prowess, to which many others will continue to appropriately attest, the most striking impact has been through the deep care and concern that he has shown me, personally, every step of the way. This characteristic is the one I am most grateful for, and is something I aspire to emulate going forward.

Finally, to my family and friends who have supported me throughout this exciting but yet very challenging chapter of my life. Your steadfast love and encouragement have been crucial at every point along the way, and for this I will be forever grateful.

For Brittany

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Chapter 1

Introduction

1.1 Tropical cyclones and climate

Tropical cyclones (TCs), referred to as hurricanes in the Atlantic, typhoons in the Pacific, and cyclones in the Indian ocean basins, respectively, are the most violent large-scale convective systems in the atmosphere. Those that make landfall are responsible for some of the most devastating of Earth's recorded natural disasters, both in their impact on human life and well-being, and also in the sheer magnitude of the associated economic losses. The Great Bhola cyclone, for example, that hit Bangladesh in 1970 and killed more than 300,000 people, counts as one of the most deadly natural catastrophes of the past century (Frank and Husain, 1971), while more recently Hurricane Katrina in 2005, in addition to the grievous human toll, caused economic losses that have been estimated well in excess of 100 billion US dollars (Knabb et al., 2005). Improving the prediction, detection and effective warning of tropical storms is appropriately one of the highest priorities of meteorological agencies around the world, along with the research community that supports them. Therefore, understanding the influence of large-scale environmental factors on TC formation has been an urgent task of great scientific and societal importance.

The past few decades have seen the emergence of anthropogenic greenhouse gas related global warming as a topic of enormous interest and concern among scientists, and in the public conscience more broadly. The relatively short period of reliable global observations of tropical cyclone activity make it very difficult to have confidence in discerning any detectable long-term trends (Knutson et al., 2010b), hence the influence of observed climate change on past TC frequency and intensity remains uncertain. However, with continuing and future changes to the climate system inevitable, both natural and anthropogenic, many scientific resources have been employed to investigate projections of future TC activity related to these changes.

Studies of future projections of TC activity in a warming world were recently summarized by the United Nations Intergovernmental Panel on Climate Change’s Fifth Assessment Report (IPCC AR5; Christensen et al., 2013), which points towards a slight increase in the frequency of intense Category 4-5 TCs and the lifetime maximum intensity, but an overall reduction in the number of global tropical storms (see Figure 1.1), with great uncertainty in the seasonal and geographic distribution of the changes. Despite a projected reduction in the frequency of tropical cyclones, Peduzzi et al. (2012) warns that mortality risk depends on several factors, including tropical cyclone intensity, exposure, and levels of poverty and governance. Peduzzi et al. (2012) concludes that projected increases in both demographic pressure and tropical cyclone intensity over the next 20 years can be expected to greatly increase the number of people exposed per year and exacerbate disaster risk, even allowing for potential progression in development and governance.

Therefore, the effect of climate change on future tropical cyclone activity continues to garner significant interest within the scientific community and beyond, and the impetus for understanding the connection between the climate and resulting TC activity has grown considerably. In what follows we will first aim to give the reader some background as to the broad categories of approaches taken to investigate the

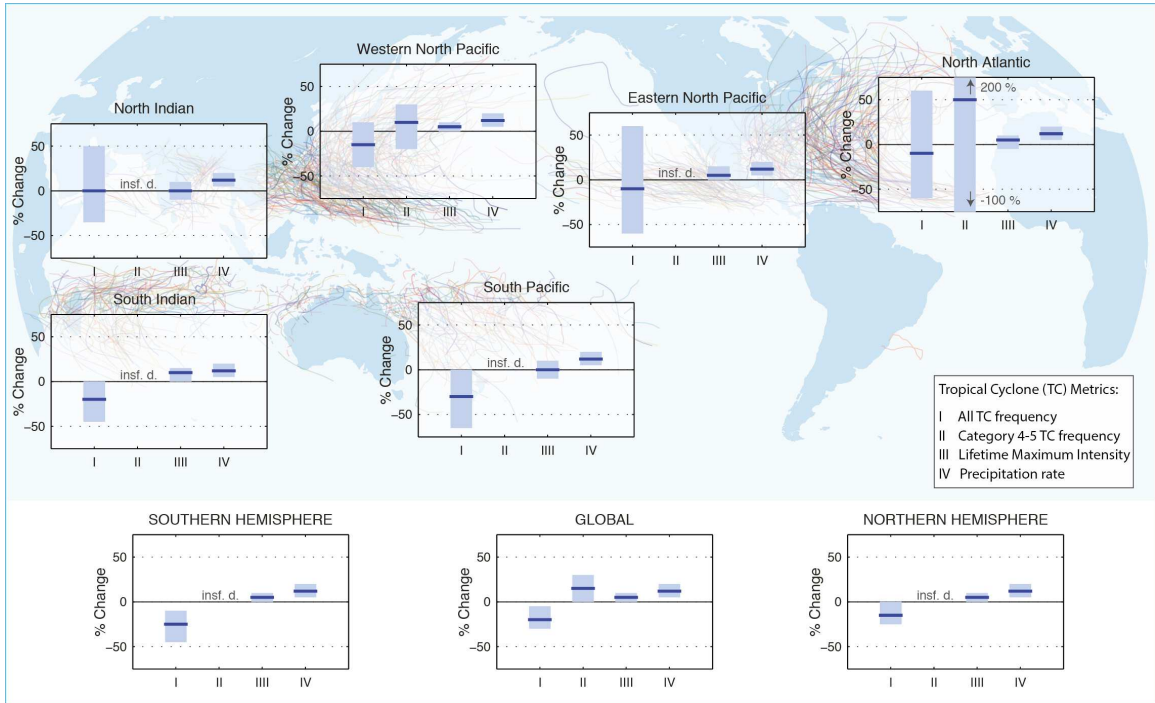


Figure 1.1: A reproduction of Fig. 14.17, Christensen et al. (2013), with permission; original caption follows: General consensus assessment of the numerical experiments described in Supplementary Material Tables 14.SM.1 to 14.SM.4. All values represent expected percent change in the average over period 2081-2100 relative to 2000-2019, under an A1B-like scenario, based on expert judgement after subjective normalization of the model projections. Four metrics were considered: the percent change in (I) the total annual frequency of tropical storms, (II) the annual frequency of Category 4 and 5 storms, (III) the mean Lifetime Maximum Intensity (LMI; the maximum intensity achieved during a storm's lifetime) and (IV) the precipitation rate within 200 km of storm centre at the time of LMI. For each metric plotted, the solid blue line is the best guess of the expected percent change, and the coloured bar provides the 67% (likely) confidence interval for this value (note that this interval ranges across -100% to $+200\%$ for the annual frequency of Category 4 and 5 storms in the North Atlantic). Where a metric is not plotted, there are insufficient data (denoted 'insf.d.') available to complete an assessment. A randomly drawn (and coloured) selection of historical storm tracks are underlain to identify regions of tropical cyclone activity.

TC-climate connection, introducing the development of so-called TC genesis indices, and the emerging opportunities provided by general circulation modeling. Some of the previous work undertaken in this endeavor will be highlighted, setting the context for the investigation conducted by the current author.

1.1.1 Relating TC formation to the large-scale environment

Meteorological observations of the thermal and dynamic structure of tropical cyclones (TCs) and their environment improved during the middle of the last century following the second world war (e.g. Simpson, 1947), allowing the re-examination and further development of theories describing TC formation (e.g. Palmén, 1948; Riehl, 1948, 1950; Yanai, 1964), although meteorologists in various parts of the world emphasized different aspects of the storm environment as being more or less important for TC development. Gray (1968) set out with the aim of developing a global climatology of tropical storm origins, convinced that there must exist similar TC development criteria across the different ocean basins. Gray’s research confirmed several basic “ingredients” - necessary conditions for TC genesis, which can be broadly stated as:

1. relatively high underlying sea surface temperature (SST),
2. sufficient ambient vorticity,
3. low vertical windshear, and
4. abundant moisture.

While much has been learned about the genesis of tropical cyclones over the subsequent decades, in large part due to the huge improvement in the coverage and quality of TC observing platforms and the increase in numerical modeling capabilities, a dynamical theory linking the annual global frequency of approximately 90 ± 10 TCs per year (Emanuel, 2008) to the large-scale environment remains elusive. In the absence of such a theory, several alternative approaches have been taken in order to interrogate the relationship between the large-scale environment and the frequency and intensity of TCs. One method builds on Gray’s ingredients-based approach and formulates empirical relationships between the observed large-scale environment and the statistics of recorded TCs. A complementary approach has sought to harness

the computational power of global circulation models that are capable of directly simulating TC-like disturbances. We now briefly introduce each of these approaches.

Tropical cyclone genesis indices

An empirical index for tropical cyclone genesis is formulated as a function of a set of predictors, being the various atmospheric fields thought to be important for enhancing or inhibiting the likelihood of TC genesis. An index that relates the frequency of genesis events to the large-scale environment is useful in several ways. Firstly, it enables a quantification of the relative importance of the parameters that comprise the index, and can thus help identify the climatological factors that inhibit or enhance the propensity for genesis in different geographic regions or seasons. Secondly, once skill in reproducing the past climatology of TC genesis has been demonstrated, an index can potentially be useful in developing schemes to predict TC frequency based on a forecast of the evolution of the large-scale environment.

Building upon his previous findings, Gray (1975, 1979) was the first to develop such an index for tropical cyclogenesis. His seasonal genesis parameter related the frequency of seasonal TC formation to six key physical parameters: three ‘dynamic potential’ parameters (relative vorticity, Coriolis and vertical wind shear), and three ‘thermal potential’ parameters (ocean thermal energy, moist static stability and relative humidity). Subsequent studies further tested (e.g., Watterson et al., 1995; Royer et al., 1998) and refined this ingredients-based approach, yielding various empirical or semi-empirical TC genesis indices based on similar environmental parameters. Most notably over the last decade has been the development and use of the Genesis Potential Index (GPI; Emanuel and Nolan, 2004; Camargo et al., 2007a,b; Nolan et al., 2007; Vecchi and Soden, 2007; Emanuel, 2010; Emanuel and Sobel, 2013), and more recently a Poisson regression index (Tippett et al., 2011; Camargo et al., 2014).

Although these different indices do a reasonable job at capturing the geographical and seasonal distribution of the observed frequency of genesis (Menkes et al., 2012; Camargo et al., 2014) in the current climate (over which the indices have been deduced), it is still far from certain whether these same relationships will apply to different climates. Consequently, along with the further refinement of this suite of genesis indices, attention is now also focused on alternative methods of projecting the potential for TC genesis that may be more applicable in different climates.

Tropical cyclones in global climate models

It was Manabe et al. (1970) who first discovered that low-resolution atmospheric general circulation models (AGCMs) were able to form tropical disturbances reminiscent of observed tropical cyclones, finding “the tropical cyclone in the model atmosphere... [developed as] a genuine consequence of the physical processes introduced into the model”, and was not merely an artifact of an error introduced from the finite differencing in the model. A decade later, Bengtsson et al. (1982) using the 1980 ECMWF operational GCM, having significantly higher horizontal resolution than the model used by Manabe et al. (1970), conducted a systematic study of numerous TC-like vortices generated by the model. While these model-generated TCs did not exhibit all of the features of observed TCs (McBride, 1984), the results were quite promising given the relatively coarse grid resolution (compared to present-day AGCMs). Especially impressive was the demonstrated ability to reproduce some of the features of the geographical and seasonal distribution of genesis locations.

Building upon this, (Krishnamurti, 1988) evaluated the track and structure of model-generated storms and investigated the impact of increasing model resolution. Early in the 1990s, Broccoli and Manabe (1990) (followed by Haarsma et al., 1993)) used a GCM coupled to a simple ocean mixed-layer model to investigate the impact of a doubling of atmospheric CO_2 on TC frequency, although initial skepticism was

leveled as to whether the GCMs had yet been proven an appropriate tool for conducting such studies (e.g. Evans, 1992). However, subsequent studies (Wu and Lau, 1992; Bengtsson et al., 1995, 1996; Royer et al., 1998) continued to demonstrate an ability to simulate TCs when forced with observed climatological SSTs, and a series of papers followed (Vitart et al., 1997, 1999; Vitart and Anderson, 2001) that utilized an ensemble of GCM integrations to investigate the impact of interannual and inter-decadal SST variability on simulated TC frequency, showing remarkable consistency with observed TCs.

Over the past decade many different modeling groups have successfully employed moderate- to high-resolution GCMs to investigate TC formation (e.g. Sugi et al., 2002; Walsh et al., 2004; Yoshimura and Sugi, 2005; Oouchi et al., 2006; Sugi et al., 2009; Zhao et al., 2009; Wehner et al., 2010; Zhao and Held, 2010; Zhao et al., 2010; Vecchi et al., 2011; Held and Zhao, 2011; Murakami et al., 2012b,a; Zhao and Held, 2012; Zhao et al., 2012; Sugi et al., 2012; Li et al., 2013; Merlis et al., 2013; Vecchi et al., 2013; Tory et al., 2013); see Shaevitz et al. (2014) for a recent review. GCMs are now routinely run at sufficient resolution to enable the internal generation of TC-like disturbances within the simulations and can thus provide an important tool for investigating the behavior of TCs in different climates.

1.1.2 Tropical cyclones and climate change

Towards the latter part of the 20th century a consensus was emerging that climate change due to increasing anthropogenic greenhouse gases would lead to increased tropical sea surface temperatures, summarized by the Intergovernmental Panel on Climate Change (IPCC; Houghton et al., 1990, 1992). Since TCs extract latent and sensible heat from the warm tropical oceans (Emanuel, 1987, 1988), the IPCC initially expressed concerns in its First Assessment Report (1990) that:

There is some evidence from model simulations and empirical considerations that the frequency per year, intensity and area of occurrence of tropical disturbances may increase [in a doubled carbon dioxide world], though it is not yet compelling. (Houghton et al., 1990)

However, in the years that followed it became clearer that any changes in tropical cyclone activity would not be solely dependent on changes in SST alone. Rather, equally important changes in moisture and stability in an enhanced CO₂ world needed to be considered. The previously-suggested absolute SST threshold for cyclogenesis ($\sim 26\text{--}27^\circ\text{C}$; Palmén, 1948; Gray, 1968), would itself also increase because of compensating changes in the tropospheric moist static stability (Emanuel, 1995). Furthermore, potential changes in the geographical pattern and magnitude of vertical wind shear, along with changes in the prevalence of necessary meteorological precursor disturbances could also considerably influence TC development. With the release of the Second Assessment Report in 1996, the IPCC expressed greater uncertainty in any projected future changes in TCs:

The formation of tropical cyclones depends not only on sea surface temperature, but also on a number of atmospheric factors. Although some models now represent tropical storms with some realism for present day climate, the state of the science does not allow assessment of future changes. (Houghton et al., 1996)

While projections of the frequency of tropical cyclone formation continued to show little or no evidence for future change (Henderson-Sellers et al., 1998), by the turn of the century in the Third Assessment Report the IPCC reported an emerging consensus from modeling studies that maximum tropical cyclone intensities (i.e. wind speeds) were “likely [to increase], in some regions” (Giorgi et al., 2001), consistent with the theoretical results of Emanuel (1987) and Holland (1997).

The increased capability over the past decade of high-resolution GCMs capable of directly simulating TCs has allowed much focused attention from the research community on probing the projected changes to TC frequency, particularly following Hurricane Katrina in 2005. Most of these studies have projected a slight reduction in the global number of tropical cyclones in response to global warming (e.g., Sugi et al., 2002; Yoshimura and Sugi, 2005; Yoshimura et al., 2006; Bengtsson et al., 2007; Emanuel, 2008; Gualdi et al., 2008; Sugi et al., 2009; Zhao et al., 2009; Held and Zhao, 2011; Murakami et al., 2012a,b), although there have been a few exceptions that have instead suggested a possible increase (e.g., Wehner et al., 2010; Emanuel, 2013); for reviews, see Knutson et al. (2010a,b, 2013) and Walsh et al. (2015).

Some studies have suggested that the decrease in TC frequency in response to global warming is connected to the reduction in the upward convective mass flux in the tropics (e.g. Sugi et al., 2002; Yoshimura et al., 2006; Bengtsson et al., 2007; Held and Zhao, 2011), which itself is a consequence of the changing large-scale hydrological cycle (e.g. Held and Soden, 2006). One interpretation of this result may be that the decreased mass flux makes it easier for the advection of dry air to suppress genesis (Nolan and Rappin, 2008; Held and Zhao, 2011).

The current consensus of a slight reduction in future TC frequency is reflected in the most recent IPCC Fifth Assessment Report (Christensen et al., 2013, recall Fig. 1.1), which also shows the partially compensating (or perhaps even overcompensating) impact of a shift to storms of slightly higher intensity, when considering changes in the number of Category 4-5 TCs and the destructive potential of TC activity (Emanuel, 2008; Bender et al., 2010). There is less confidence in future projections when looking at individual basins as to the sign or magnitude of the change (Knutson et al., 2010b; Christensen et al., 2013), or when considering only the subset of storms that attain a certain threshold intensity. The spread in the range of responses

may reflect uncertainty arising from different large-scale forcing (Villarini et al., 2011; Villarini and Vecchi, 2013; Knutson et al., 2013).

Among other factors, studies have shown a strong correlation between projected changes in tropical cyclone activity and the spatial pattern of tropical SST (e.g. Vecchi and Soden, 2007; Vecchi et al., 2008; Sugi et al., 2009; Murakami et al., 2012a; Zhao and Held, 2012; Camargo et al., 2014). Hence, projections of regional tropical cyclone activity will depend critically on the reliability of projections of the change in SST patterns. The current study will explore the impact of changes in the meridional and zonal pattern of SST on TC activity within an idealized GCM.

1.2 TC activity in an aquaplanet GCM

We have been accustomed for several decades to the use of idealized and comprehensive global atmospheric models in tandem to study the controls on *midlatitude* eddies (e.g., Brayshaw et al., 2008; Lu et al., 2010), justified by the quality of the extratropical simulations provided by the global models. Given the demonstrated ability of current global climate models to simulate TCs (Walsh et al., 2007; Shaevitz et al., 2014), we believe analogous idealization and manipulation is warranted, toward the goal of better understanding the large-scale controls on tropical cyclogenesis.

The current study follows an extensive series of experiments using a 50-km horizontal resolution global atmospheric GCM to look of the response of simulated TCs to various changes in Earth’s climate (e.g., Zhao et al., 2009, 2010; Held and Zhao, 2011; Zhao and Held, 2012; Vecchi et al., 2013; Merlis et al., 2013; Camargo et al., 2014); the model will be introduced in greater detail in the chapter that follows. To remove complexities arising from the geometric distribution of the continents we employ an aquaplanet configuration, and design a set of new experiments to investigate the response of simulated tropical cyclones to systematic variations in the underlying

SST pattern and the associated large-scale circulation. To begin with, the focus will be on zonally-symmetric climates, controlled by the zonal-mean meridional SST profile. Later we introduce zonally-asymmetric perturbations to explore the impact of changes to the zonal pattern of SST. Some of the results from this study are already published in Ballinger et al. (2015).

1.2.1 Thesis outline

We begin in Chapter 2 with an introduction of the aquaplanet model that has been used throughout this study, and show the formulation of the lower boundary conditions that are applied to the suite of experiments that follow. The automated system for detecting and tracking TC-like vortices within the model output fields is also discussed, as well as our chosen intensity thresholds for classifying TCs within the study.

In Chapter 3 we begin by investigating the sensitivity of tropical cyclone activity to changes in the latitude of the maximum SST, while holding the magnitude and gradient of the zonally-symmetric SST profiles fixed. Following this we apply warming and cooling SST perturbations to the lower boundary condition, firstly as a simple globally-uniform perturbation to explore the response of TCs to uniform warming or cooling of the SSTs. The next suite of experiments confine the warm or cool perturbation to the tropical region, centered on the latitude of the maximum SST. The main dynamical effect comes from the meridional gradient of SST. In summary, the third chapter seeks to carefully investigate changes in TC activity resulting from:

- changes in the **location** of the maximum SST,
- changes in the **magnitude** of global-mean SST perturbation, and
- changes in the tropical **meridional gradient** of SST.

In Chapter 4 the results of the third chapter are extended with the aim of investigating the evolution of tropical cyclone intensity within the model following genesis. We introduce a Lagrangian framework by which changes in the mean characteristics of the storms can be related to the surrounding atmospheric conditions. In Chapter 5 we analyze a new set of experiments with a zonally-asymmetric boundary condition, designed to explore the sensitivity of TC frequency to changes in the tropical **zonal gradient** of SST. We summarize the findings from this study in Chapter 6.

Chapter 2

The HiRAM Aquaplanet

The Geophysical Fluid Dynamics Laboratory High-Resolution Atmospheric Model (HiRAM) uses a cubed-sphere atmospheric dynamical core with approximately 50 km horizontal resolution and 32 vertical levels (Zhao et al., 2009). When observed monthly-mean sea surface temperatures are prescribed as the lower boundary condition, HiRAM is able to simulate the seasonal cycle of tropical cyclone frequency in individual ocean basins (Zhao et al., 2009). In addition, HiRAM captures the inter-annual variability in Atlantic hurricane frequency, has seasonal forecast skill, and has been used for projections of future tropical cyclone frequency (e.g., Zhao et al., 2009, 2010; Zhao and Held, 2012; Camargo et al., 2014). In a recent comparison study, Shaevitz et al. (2014) identifies HiRAM as simulating the TC climatology remarkably well across a variety of metrics.

A substantial fraction of the tropical precipitation in HiRAM occurs in resolved motion; however, there is a convection parameterization, based on a bulk entraining-detraining plume (Bretherton et al., 2004), and the top-of-atmosphere radiation balance and tropical cyclone frequency are sensitive to the parameterization’s entrainment parameter (Zhao et al., 2009, 2012). The aquaplanet simulations that are presented use identical sub-grid scale parameterizations to the HiRAM simulations with

comprehensive boundary conditions. HiRAM simulations are presented in two different aquaplanet configurations: one with a fixed, zonally symmetric SST boundary condition and the other with interactive SSTs determined by a slab ocean boundary condition.

2.1 Fixed SST experiments

In the fixed-SST aquaplanet HiRAM simulations, we prescribe a zonally symmetric surface temperature profile of the form

$$SST(\theta) = \begin{cases} SST_0 \left[1 - \frac{1}{2} \left(\sin^2 \left(\frac{\theta - \theta_0}{140} \pi \right) + \sin^4 \left(\frac{\theta - \theta_0}{140} \pi \right) \right) \right] & ; \quad \theta_0 - 70^\circ < \theta < \theta_0 + 70^\circ \\ 0 & ; \quad \text{elsewhere,} \end{cases} \quad (2.1)$$

where SST ($^\circ\text{C}$) is the surface temperature as a function of latitude θ (degrees). The functional form of $SST(\theta)$ is similar to the *Qobs* profile proposed in the Neale and Hoskins (2001) suite of aquaplanet experiments, and is chosen here to approximate Earth’s observed zonal-mean SST distribution. The maximum temperature SST_0 occurs at latitude θ_0 (i.e., $SST(\theta_0) = SST_0$), and the profile decreases symmetrically to latitudes $\theta_0 \pm 70^\circ$, beyond which the SST is fixed at 0°C .

A series of three perpetual-summer control simulations were conducted in which the latitude of the SST maximum was varied by setting $\theta_0 = \{10, 13, 16\}$ degrees, all with $SST_0 = 28.5^\circ\text{C}$. HiRAM simulations where the prescribed SST maximum is equatorward of $\sim\pm 10^\circ$ (not shown here) do not generate a sufficient number of TCs for the analysis. However, as will be shown in the following section, the frequency of cyclones varies markedly as the maximum in SST latitude shifts from 10°N to 16°N , and hence the chosen latitudes of the control experiments allow a focus on the sensitivity of these changes.

Perturbation SST profiles are then formed by adding and subtracting a temperature perturbation SST' to the control profiles, $SST_{\pm}(\theta) = SST(\theta) \pm SST'(\theta)$.

The first set of SST perturbation simulations were designed to investigate the sensitivity of TC activity to a globally-uniform warming or cooling. A uniform 1.5°C was added to the SSTs of the furthest poleward control run ($\theta_0 = 16^{\circ}\text{N}$) to form the global warming (SST_{G+}) profile. Similarly, a uniform 1.5°C was subtracted from the $\theta_0 = 16^{\circ}\text{N}$ control run to form the global cooling (SST_{G-}) profile.

The second set of SST perturbation simulations was designed to investigate the sensitivity of TC activity to a warming or cooling confined to tropical latitudes. The prescribed tropical perturbation has the form

$$SST'(\theta) = \begin{cases} SST'_0 \left[1 - \frac{1}{2} \left(\sin^2 \left(\frac{\theta - \theta_0}{60} \pi \right) + \sin^4 \left(\frac{\theta - \theta_0}{60} \pi \right) \right) \right] & ; \quad \theta_0 - 30^{\circ} < \theta < \theta_0 + 30^{\circ} \\ 0 & ; \quad \text{elsewhere,} \end{cases} \quad (2.2)$$

where SST'_0 is the magnitude of the perturbation centered on the latitude of the SST maximum (in the control run), with the perturbation decaying to zero at $\theta_0 \pm 30^{\circ}$. Since the meridional extent of the surface temperature perturbation is narrower ($\theta_0 \pm 30^{\circ}$) than that of the control profiles ($\theta_0 \pm 70^{\circ}$), the resulting warmer ($SST_{T+} = SST + SST'$) and cooler ($SST_{T-} = SST - SST'$) tropical profiles have a corresponding sharper or flatter meridional tropical SST gradient, respectively. Six additional simulations were performed by setting $SST'_0 = 1.5^{\circ}\text{C}$, tropical perturbation simulations (warmer and cooler) were conducted for each of the three control profiles, six additional simulations in total.

For each simulation the top-of-atmosphere incoming solar radiation is set at a perpetual equinox, and the ozone distribution and greenhouse gas concentrations are hemispherically symmetric following similar aquaplanet experiment conventions (e.g., Neale and Hoskins, 2001). The only aspect of the simulations that is hemispherically

Configuration	Experiment	Length of simulation
Fixed SSTs	$\theta_0 = 10^\circ \text{ N}$ (control)	10 yr
	$\theta_0 = 13^\circ \text{ N}$ (control)	10 yr
	$\theta_0 = 16^\circ \text{ N}$ (control)	10 yr
	$\theta_0 = 10^\circ \text{ N}$ (+1.5° C tropics)	5 yr
	$\theta_0 = 13^\circ \text{ N}$ (+1.5° C tropics)	5 yr
	$\theta_0 = 16^\circ \text{ N}$ (+1.5° C tropics)	5 yr
	$\theta_0 = 10^\circ \text{ N}$ (−1.5° C tropics)	5 yr
	$\theta_0 = 13^\circ \text{ N}$ (−1.5° C tropics)	5 yr
	$\theta_0 = 16^\circ \text{ N}$ (−1.5° C tropics)	5 yr
	$\theta_0 = 16^\circ \text{ N}$ (+1.5° C uniform)	5 yr
	$\theta_0 = 16^\circ \text{ N}$ (−1.5° C uniform)	5 yr

Table 2.1: A summary of the HiRAM aquaplanet fixed-SST simulations

asymmetric is the prescribed SST. Each control simulation was run with its prescribed SST profile for a period of 10 yr (10×365 model days); the perturbation simulations were each run for a period of 5 yr (Table 2.1).

2.2 Slab ocean experiments

In the HiRAM slab ocean simulations, which are the same as in Merlis et al. (2013), the SST evolves in response to the turbulent surface enthalpy fluxes, surface radiative fluxes and meridional heat flux in an energetically consistent fashion. The heat capacity of the surface is equivalent to a water depth of 20 m. The surface albedo is uniformly 0.08, and there is no representation of sea ice. Hence, the surface temperature can be lower than the freezing point of water, which occurs in the high latitudes of these simulations. In addition to the surface fluxes and radiation, we prescribe an ocean heat flux convergence in the slab ocean. This is the only aspect of the model formulation that is hemispherically asymmetric.

The design of the prescribed ocean heat flux convergence Q in the slab ocean experiments follows Kang et al. (2008). A surface heating is applied to the northern hemisphere extratropics (poleward of 40°), and a compensating surface cooling is

applied to the southern hemisphere extratropics of the form

$$Q(\theta) = \begin{cases} Q_0 \sin\left(\frac{\theta + 40}{50}\pi\right) & ; \quad -90^\circ < \theta < -40^\circ \\ 0 & ; \quad -40^\circ \leq \theta \leq 40^\circ \\ Q_0 \sin\left(\frac{\theta - 40}{50}\pi\right) & ; \quad 40^\circ < \theta < 90^\circ, \end{cases}$$

where Q_0 is the amplitude of the heat flux convergence in Wm^{-2} . Because the pattern and magnitude of the heating and cooling are equal and opposite, no heat is added or subtracted from the global system. Thus the forcing in these experiments is equivalent to an implied cross-equatorial heat flux F_0 , where $Q = -\nabla \cdot F_0$ [see Fig. 1 of Kang et al. (2008)].

The insolation is an idealized function of latitude $S_{\downarrow} = \tilde{S}_0 \times \widetilde{\cos \zeta}$ with $\tilde{S}_0 = S_0/2$ and

$$\widetilde{\cos \zeta} = 2 \left[1 + \frac{\Delta_s}{4} (1 - \sin^2 \theta) \right],$$

where $S_0 = 1400 \text{ W m}^{-2}$ and $\Delta_s = 0.9$. The factor of two increase to the cosine of the zenith angle corresponds to a day-time average of the diurnal cycle (see Cronin, 2014, for other choices), and is approximately an insolation-weighted annual-mean of the day-time average cosine of the zenith angle for orbital parameters similar to Earth's. With $\Delta_s = 0.9$, the insolation gradient is somewhat stronger than Earth's annual-mean, but is weaker than equinox conditions (the insolation is not zero at the pole). The CO_2 concentration is 300 ppm and there are no other greenhouse gases or aerosols. The ozone distribution (varying with height and latitude) is fixed as in Neale and Hoskins (2001), and is similar to Earth's present-day zonal-mean climatology.

A series of four simulations were performed in which the strength of the cross-equatorial heat flux was varied by setting $Q_0 = \{20, 40, 50, 60\} \text{ Wm}^{-2}$. Each simulation was run for 10 yr (see Table 2.2) from an Earth-like initial condition, with the final 5 yr retained for analyses. The 5-yr model spinup is adequate to reach a

Configuration	Experiment	Length of simulation
Slab Ocean	$Q_0 = 20 \text{ Wm}^{-2}$	10 yr
	$Q_0 = 40 \text{ Wm}^{-2}$	10 yr
	$Q_0 = 50 \text{ Wm}^{-2}$	10 yr
	$Q_0 = 60 \text{ Wm}^{-2}$	10 yr
Fixed SSTs	Slab-ocean SSTs (40 Wm^{-2})	5 yr
	Slab-ocean SSTs (60 Wm^{-2})	5 yr

Table 2.2: A summary of the HiRAM aquaplanet slab-ocean simulations

statistical steady state, and, given the absence of a seasonal cycle, 5-yr averages are sufficient for sampling tropical cyclone statistics for our purposes.

2.3 Cyclone detection and tracking

The cyclone detection and tracking algorithm is based on Vitart et al. (1997) and Knutson et al. (2007), and is discussed in detail in Appendix B of Zhao et al. (2009); it has recently been compared with other similar detection and tracking algorithms in Horn et al. (2014). First, candidate warm-core vortices are identified using 6-hourly instantaneous 850-hPa relative vorticity, sea-level pressure and a measure of the upper-tropospheric temperature anomaly. The maximum “surface” wind speed is recorded at each 6-hr output interval, being the maximum of 10-m wind speed values of gridpoints surrounding the location of the candidate vortex, $\mathbf{v}_{\text{sfc}} = \max(|\tilde{\mathbf{v}}_{10\text{m}}|_i)$. A simple tracking scheme subsequently links warm-core vortices that are within 400 km of those in the preceding 6 hr. Storms are not included in the final analysis unless their associated track lasts in excess of 3 days, with the additional requirement that a tropical storm wind speed criterion is satisfied for at least 3 days (not necessarily consecutive).

The maximum surface wind speed threshold for classifying storms (17 m s^{-1} for tropical storms, 33 m s^{-1} for hurricanes) is reduced by $\sim 10\%$ as described in Zhao

et al. (2012), which is consistent with the range recommended by Walsh et al. (2007) in order to account for the impact of the model’s ~ 50 -km spatial resolution.

All tropical storm tracks meeting the above criteria (with $\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$) are herein classified broadly as tropical cyclones (TCs), with “genesis” being defined to occur at the time and location where the storm is first analyzed having $\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$. Reference will on occasion be made to a *strong* subset of TCs that at some point during their lifetime attain hurricane-strength wind speeds ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$).

Chapter 3

Zonally symmetric forcing

3.1 General features of the simulations

Forced with hemispherically asymmetric SSTs or ocean heat flux convergence, the simulated tropical climate has a relatively narrow region of ascent and enhanced precipitation shifted off the equator in the warmer northern hemisphere, which is similar to the summer climate that is realized over many of Earth’s tropical oceanic regions. Figure 3.1 shows a snapshot of the HiRAM aquaplanet’s tropics, taken from the control fixed-SST simulation where the maximum in SST is located at 16° N. The zonally elongated band of daily precipitation rate shows the location of the model’s ITCZ, which in the time mean resides at 8° N (the latitude of maximum precipitation rate) in this simulation (left panel of Fig. 3.1).

The ITCZ location is associated with low-level convergence in the meridional wind velocity field (red line in Fig. 3.1), where the surface branches of the summer and winter (cross-equatorial) Hadley cells meet. The northward surface flow accelerates across the equator and turns to the right (eastward) under the influence of the Coriolis force. Hence the SH south-easterly trade winds and NH north-easterly trade winds form a zone of ‘monsoon’ westerlies that coincide with the off-equatorial region of

ascent (black line in Fig. 3.1). On the poleward flank of the ITCZ region the meridional gradient of the time-mean zonal-mean zonal wind is therefore negative, yielding a ‘sweet-spot’ region of enhanced low-level cyclonic vorticity coincident with surface convergence and large-scale ascent. Recall that the maximum in SST is also to the north of the ITCZ, and it is unsurprising that this region is observed to be favorable for the formation of TCs in the model.

Numerous TCs disturbances develop in HiRAM, some of which subsequently strengthen and propagate toward higher latitudes. Examples of these phenomena can be seen in Figure 3.1 in the instantaneous 10-m wind speed field, giving a sense of the range of sizes and the spatial distribution of the disturbances. The following results focus on the relationship of the frequency of simulated TCs forming on the poleward flank of the ITCZ to the large-scale environmental conditions of the region. Table 3.1 provides a summary of the various HiRAM aquaplanet simulations conducted, and the resulting mean genesis frequency of the TCs that were detected and tracked within each simulation. These results will be discussed through the remainder of this chapter.

3.2 Poleward-shift experiments

The meridional SST profiles for the three fixed-SST simulations are shown in Figure 3.2a. By construction the maximum temperature (28.5° C) does not change, but the latitude of the maximum shifts poleward in 3° increments (10° N, 13° N, and 16° N). From the location of the maximum, there is a symmetrical decrease (northward and southward) of SST with latitude ($\propto -1/2(\sin^2 \theta + \sin^4 \theta)$). For comparison, the zonal-mean of Earth’s climatological September SST (HadISST, Rayner et al., 2003) is also plotted (black dashed line). The magnitude and general shape of the prescribed SST is similar to late boreal summer conditions on Earth.

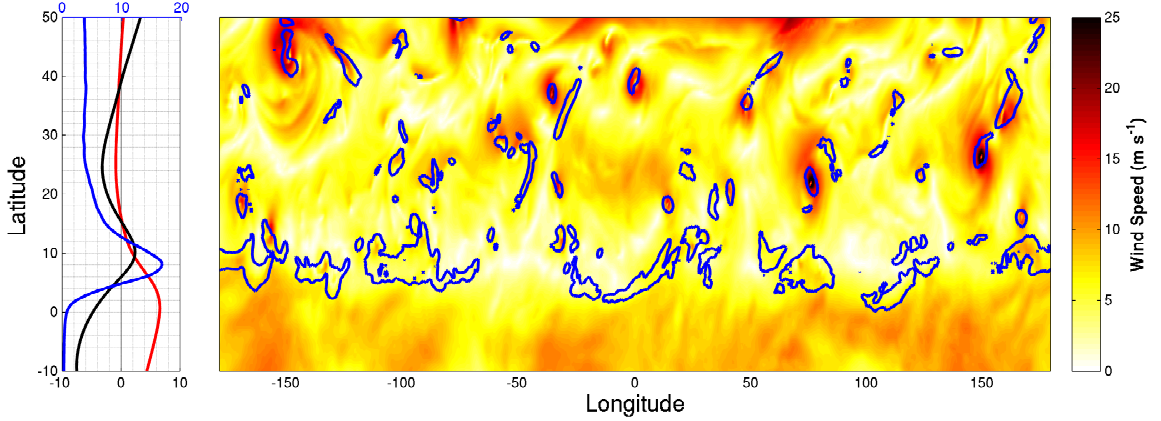


Figure 3.1: A snapshot of the 16° N control HiRAM aquaplanet simulation, displaying the instantaneous 10 m wind speed (shading, m s^{-1}) and daily precipitation rate (blue contour of 15 mm day^{-1}). Time-mean latitudinal profiles from the simulation are shown to the left: precipitation rate (blue, mm day^{-1} , scale on upper axis); and the zonal (black) and meridional (red) components of the 10-m wind (m s^{-1} , scale on lower axis).

The equilibrium SST profiles from the four slab ocean simulations are shown in Figure 3.3a. Without a cross-equatorial heat flux, the profiles would be symmetric about the equator; here the SSTs become more asymmetric as the heat flux is increased. The maximum temperature moves $\sim 2^\circ$ north for each 10 Wm^{-2} increase in heat flux amplitude. In addition, the maximum temperature itself increases by $\sim 1^\circ \text{ C}$ per 10 Wm^{-2} , but remains cooler than the maximum prescribed in the fixed-SST control runs (28.5° C) in all but the strongest heat flux slab ocean simulation. The meridional SST gradient in the slab ocean simulations is markedly sharper than those of the fixed-SST control simulations.

Figure 3.2b shows that the ITCZ (as indicated by the location of the maximum precipitation rate) moves northward when the SST maximum is shifted northward in the fixed-SST simulations. In all cases the maximum precipitation rate is equatorward of the maximum SST. A small secondary local maximum in precipitation rate to the north of the SST maximum is apparent in the 10° N experiment, but not in the experiments with the SST maximum further north. As the precipitation shifts

Configuration	Experiment	Frequency (yr^{-1})	
		<i>All TCs</i>	<i>Strong TCs</i>
Fixed SSTs	$\theta_0 = 10^\circ \text{ N}$ (control)	116.8 ± 2.1	53.8 ± 1.3
	$\theta_0 = 13^\circ \text{ N}$ (control)	183.2 ± 3.0	111.6 ± 1.5
	$\theta_0 = 16^\circ \text{ N}$ (control)	270.7 ± 3.8	173.7 ± 2.1
	$\theta_0 = 10^\circ \text{ N}$ (+1.5° C tropics)	131.3 ± 4.8	49.7 ± 2.4
	$\theta_0 = 13^\circ \text{ N}$ (+1.5° C tropics)	208.2 ± 4.2	105.2 ± 3.6
	$\theta_0 = 16^\circ \text{ N}$ (+1.5° C tropics)	317.2 ± 5.5	157.1 ± 3.8
	$\theta_0 = 10^\circ \text{ N}$ (−1.5° C tropics)	96.2 ± 4.9	53.7 ± 3.5
	$\theta_0 = 13^\circ \text{ N}$ (−1.5° C tropics)	130.3 ± 5.0	91.6 ± 2.3
	$\theta_0 = 16^\circ \text{ N}$ (−1.5° C tropics)	203.6 ± 2.8	165.3 ± 2.8
	$\theta_0 = 16^\circ \text{ N}$ (+1.5° C uniform)	222.5 ± 5.1	140.7 ± 5.2
Slab Ocean	$Q_0 = 20 \text{ Wm}^{-2}$	37.6 ± 6.6	0.4 ± 0.3
	$Q_0 = 40 \text{ Wm}^{-2}$	306.1 ± 12.5	33.0 ± 2.6
	$Q_0 = 50 \text{ Wm}^{-2}$	389.3 ± 7.1	84.7 ± 1.8
	$Q_0 = 60 \text{ Wm}^{-2}$	436.7 ± 4.6	157.3 ± 4.0
Fixed SSTs	Slab ocean SSTs (40 Wm^{-2})	361.6 ± 8.9	29.4 ± 4.0
	Slab ocean SSTs (60 Wm^{-2})	406.2 ± 7.0	166.8 ± 3.0

Table 3.1: A summary of the HiRAM aquaplanet simulations, and the resulting mean genesis frequency of the TCs tracked within each simulation. The frequency is the rate of formation equatorward of 25° N for all TCs ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ ms}^{-1}$) and strong TCs ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ ms}^{-1}$), with the mean and standard error computed over the length of the simulation. Only the final 5 yr of each of the simulations in the slab ocean configuration were included in the analysis.

poleward, it broadens meridionally and the maximum precipitation rate decreases within the ITCZ region. The broadening is more pronounced on the poleward side of the maximum, so the sub-tropics (near 15° N) receive slightly higher rainfall.

The slab ocean simulations have a similar SST-ITCZ relationship (Figure 3.3a,b). As the cross-equatorial heat flux increases the precipitation rate maximum shifts to the north and the meridional profile of the precipitation rate broadens. Here the shift is quite marked and the latitude of the maximum precipitation is closer to (though still equatorward of) the location of the maximum SST. The mean meridional circulation is generally stronger in the slab ocean simulations (not shown here), which gives rise

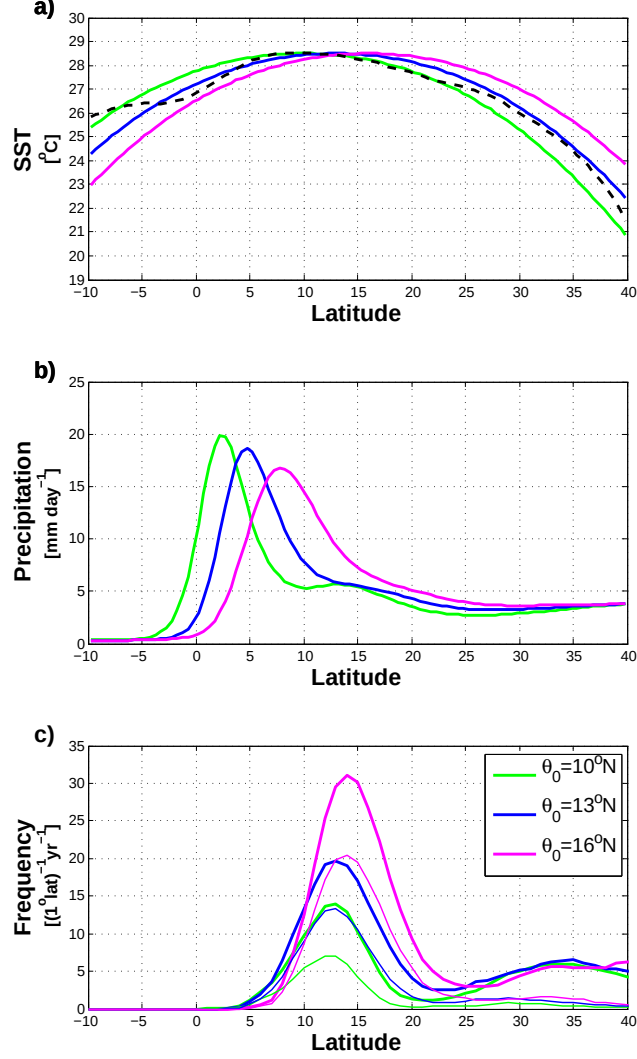


Figure 3.2: Time- and zonal-mean fields from the HiRAM aquaplanet fixed-SST control simulations: a) SST ($^{\circ}\text{C}$); b) precipitation rate (mm day^{-1}); and c) genesis frequency (TCs $(1^{\circ}\text{lat})^{-1}\text{yr}^{-1}$). The distribution of genesis frequency is shown for all tropical cyclones ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$, thick lines) and for the subset that attain hurricane-strength wind speeds ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$, thin lines). The black dashed line in the upper panel shows Earth’s zonal-mean SST during September.

to higher precipitation in the tropical convergence zone and lower precipitation in the regions of subsidence compared to the fixed-SST simulations.

The extent to which the fixed-SST simulations can reproduce the behavior of the slab ocean model can be investigated if the prescribed SSTs are taken from the climatology of the latter. This comparison was made for two of the slab ocean cases,

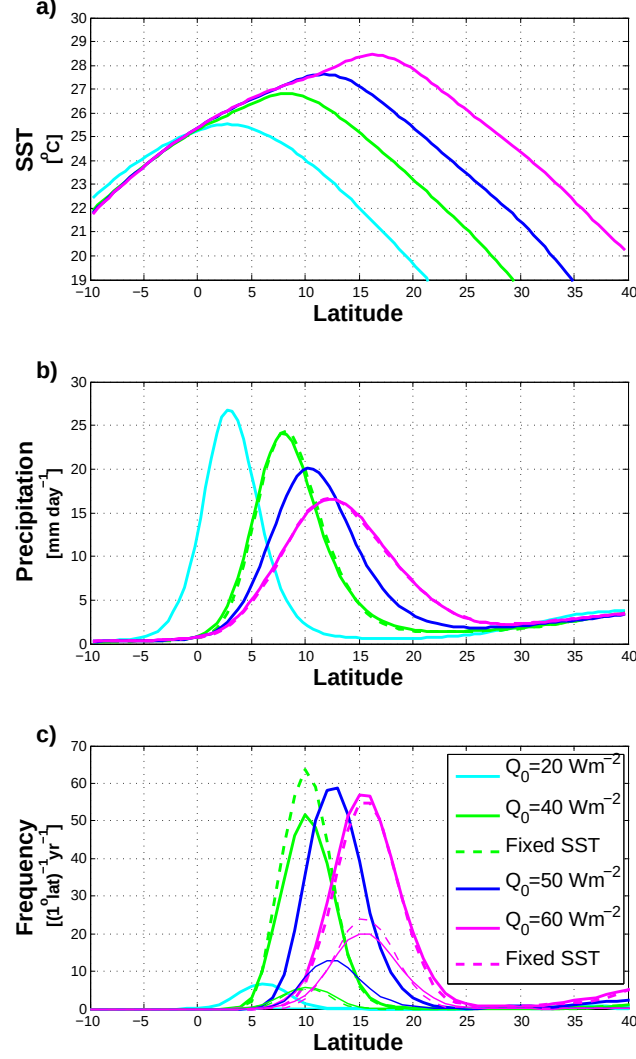


Figure 3.3: Time- and zonal-mean fields from the HiRAM aquaplanet simulations in the slab ocean configuration: a) SST ($^{\circ}\text{C}$); b) precipitation rate (mm day^{-1}); and c) genesis frequency (TCs $(1^{\circ}\text{lat})^{-1}\text{yr}^{-1}$). The distribution of genesis frequency is shown for all tropical cyclones ($v_{\text{sfc}} > 15.2 \text{ m s}^{-1}$, thick lines) and for the subset that attain hurricane-strength wind speeds ($v_{\text{sfc}} > 29.5 \text{ m s}^{-1}$, thin lines).

$Q_0 = \{40, 60\} \text{ Wm}^{-2}$, with 5-yr fixed-SST simulations with the time-mean zonal-mean SSTs previously diagnosed from the slab ocean model prescribed. The resulting zonal-mean profiles of precipitation are shown by the dashed lines in Figure 3.3b, and are barely distinguishable from their solid counterparts. Hence, the time-mean precipitation in the ITCZ region is (to a large extent) reproduced when the slab

ocean lower boundary condition is replaced by fixed SSTs that have been fixed to the climatological average.

Meridional profiles of TC genesis frequency are produced from the results of the tropical cyclone tracking scheme (Sec. 2.3). For each simulation, the TC genesis locations are sorted into 1° latitude bins and the resulting histogram is normalized (TCs $(1^\circ \text{ lat})^{-1} \text{ yr}^{-1}$) and smoothed using a 5° latitude moving-average for display purposes. The resulting profiles of genesis frequency for the fixed-SST simulations are shown in Figure 3.2c. Thicker lines show genesis profiles of all TCs ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$) while thinner lines denote the subset of these storms that at some stage during their lifetime attain hurricane-strength wind speeds ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$).

Genesis primarily occurs on the poleward flank of the ITCZ, and there is little qualitative difference in the latitudinal distribution of genesis when comparing all TCs and the stronger hurricane subset. While there is a slight northward shift in genesis as the SST and ITCZ moves northward, the primary genesis region remains within the $\sim 8^\circ \text{ N}$ to $\sim 18^\circ \text{ N}$ latitude range in these experiments. However, as the maximum SST shifts north and the ITCZ shifts and broadens, a substantial increase in genesis frequency is simulated. With the maximum SST located at 10° N , fewer than 15 TCs per year form in the most favorable (1° latitude) genesis latitude. This frequency increases to more than 30 TCs per year when the maximum SST is located further poleward at 16° N . The proportion of storms that achieve hurricane-strength wind speeds increases slightly from $\sim 45\%$ (10° N simulation) to $\sim 65\%$ (16° N simulation).

The meridional profiles of TC genesis frequency for the slab ocean simulations are shown in Figure 3.3c. There is a marked increase in TC genesis frequency as the SST maximum and ITCZ shift poleward, as in the fixed-SST simulations. There is a substantial increase in the frequency of genesis from the simulation with the weakest ocean heat transport experiment ($Q_0 = 20 \text{ W m}^{-2}$, light-blue line) to the next simulation in the series (the heat flux convergence is doubled to $Q_0 = 40 \text{ W m}^{-2}$, green

line). Further increases in TC genesis frequency occur as the asymmetry increases, although the rate of increase slows and there are indications of a potential saturation in the number of TCs developing in the model. As the asymmetry is increased, the genesis region moves northward at a rate similar to the shift in the profile of mean precipitation, remaining on the poleward flank of the ITCZ and broadening slightly. Very few of the TCs that form in the weakest experiment reach hurricane-strength (2 out of 188 TC tracks in the 5 yr analyzed), but this proportion increases to $\sim 36\%$ in the $Q_0 = 60 \text{ Wm}^{-2}$ simulation.

Some interesting changes emerge in the resulting frequency of TCs when the simulations are conducted holding the SSTs fixed to the slab (Fig. 3.3c dashed lines, and see also the bottom two rows of Table 3.1). The $Q_0 = 40 \text{ Wm}^{-2}$ comparison yields a greater number of TCs in the fixed-SST configuration, whereas the $Q_0 = 60 \text{ Wm}^{-2}$ comparison yields fewer TCs. However, there are also changes in the intensity distribution of the resulting TCs: slightly fewer achieve hurricane-strength in the $Q_0 = 40 \text{ Wm}^{-2}$ case with fixed-SSTs, while a greater number of TCs reach hurricane-strength in the fixed-SST $Q_0 = 60 \text{ Wm}^{-2}$ simulation than the corresponding slab ocean simulation. These differences warrant further investigation, but the increases in TC genesis frequency in response to increasing hemispheric asymmetry holds in both the fixed-SST and slab ocean configurations.

3.3 Global perturbation experiments

In order to test the sensitivity of genesis frequency to the magnitude of the SSTs, fixed-SST globally-uniform perturbation experiments were conducted as outlined earlier (Sec. 2.1). Results from the 16° N runs are shown in Figure 3.4. The control run (black line) is the same as the 16° N profile that was plotted in Figure 3.2; also shown

here are the meridional profiles associated with the warm and cool global perturbation experiments, plotted in red and blue respectively.

Figure 3.4 shows results from a globally-uniform 1.5°C perturbation. By construction the meridional gradient of SST remains fixed in these runs (Fig. 3.4a). The maximum in the time-mean precipitation rate (Fig. 3.4b) increases in the warmer-SST simulation, consistent with various previous theoretical and modeling studies that describe changes to the hydrological cycle in a warmer climate (e.g., Allen and Ingram, 2002). The model’s ITCZ also moves slightly equatorward in the warmer climate. In the globally cooler SST scenario the precipitation rate reduces slightly, with no clear shift in the position of the ITCZ.

The resulting TC genesis frequency is shown in Figure 3.4c. A reduction (increase) in the genesis frequency of TCs occurs in the simulations with a globally warmer (cooler) SST. A similar change occurs in the subset of stronger TCs, such that the proportion ($\sim 65\%$) of hurricane-strength TCs is similar in both the SST_{G+} and SST_{G-} simulations. Table 3.1 shows that the simulated changes in genesis frequency (of all TCs) is -17.8% for a 1.5°C uniform warming and $+5\%$ for a 1.5°C uniform cooling. These results are consistent with previous studies by Yoshimura and Sugi (2005) and Held and Zhao (2011) who found a reduction in global genesis frequency (-12.4% and -11.0% , respectively) with a 2°C uniform warming in SST, albeit with a smaller percentage decrease that may result from more realistic Earth-like boundary conditions (e.g., seasonally varying SSTs, and including continents). The uniform SST warming and cooling effects on tropical cyclone frequency appear to be nonlinear, a result also consistent with Yoshimura and Sugi (2005) who found no change in frequency in their 2°C uniform cooling simulation.

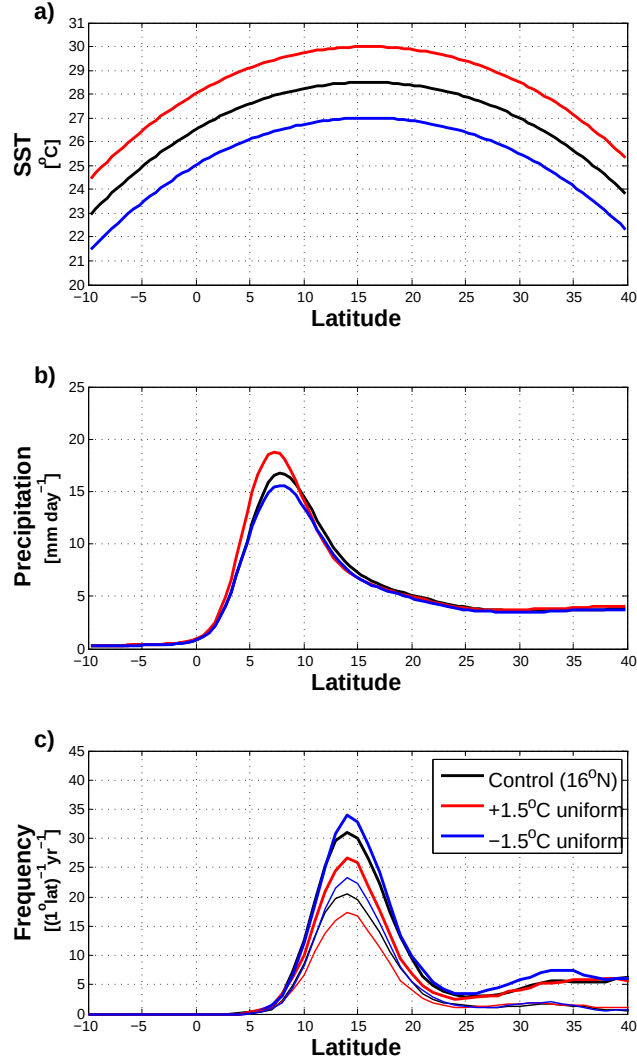


Figure 3.4: Time- and zonal-mean fields from the HiRAM aquaplanet fixed-SST simulations with a globally-uniform SST perturbation: a) SST ($^{\circ}\text{C}$); b) precipitation rate (mm day^{-1}); and c) genesis frequency (TCs $(1^{\circ}\text{lat})^{-1}\text{yr}^{-1}$). The distribution of genesis frequency is shown for all tropical cyclones ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$, thick lines) and for the subset that attain hurricane-strength wind speeds ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$, thin lines).

3.4 Tropical perturbation experiments

In order to test the sensitivity of genesis frequency to changes in the tropical SST gradient, fixed-SST tropical perturbation experiments were conducted as outlined earlier (Sec. 2.1). Results from the 16°N runs are shown in Figure 3.5. The control run (black line) is the same as the 16°N profile that was plotted in Figure 3.2; also

shown here are the meridional profiles associated with the warm and cool tropical perturbation experiments, plotted in red and blue respectively.

The results of simulations with a 1.5°C tropical perturbation that is centered on the maximum SST and reduces to zero at $16^{\circ}\text{N} \pm 30^{\circ}$ are shown in Figure 3.5. Note the steeper (flatter) meridional SST gradients associated with the warm (cool) tropical perturbation (Fig. 3.5a). As in the globally-uniform perturbation, the warmer tropical SST simulation has a higher average precipitation rate (Fig. 3.5b). However, the tropical warming SST perturbation simulation has a marked poleward shift ($\sim 2^{\circ}$) of the ITCZ toward the latitude of the maximum SST. In the case of the tropical cooling perturbation, there is a slight reduction in the maximum precipitation rate and a substantial equatorward shift ($\sim 3^{\circ}$) of the ITCZ.

The genesis statistics in Figure 3.5c show a substantial increase (reduction) in the genesis frequency of all TCs in the simulations that have warmer (cooler) tropical SSTs. That is, the simulations with sharper (flatter) meridional SST gradients in the tropics yield more (fewer) TCs. While there is a marked increase and decrease in the amplitude of the genesis frequency distributions (compared with the control), there are only slight changes in the location of the peak of the distributions and the width of the latitude bands in which genesis occurs. The results differ when focusing on the subset of stronger TCs. About 50% of TCs attain hurricane-strength wind speeds in the SST_{T+} simulation, compared to more than 80% in the SST_{T-} simulation. Hence, the number of hurricane-strength TCs changes less than the total number of TCs because of opposing changes in the intensity distribution: TCs in simulations with sharper (flatter) meridional SST gradients are less (more) likely to reach hurricane-strength wind speed.

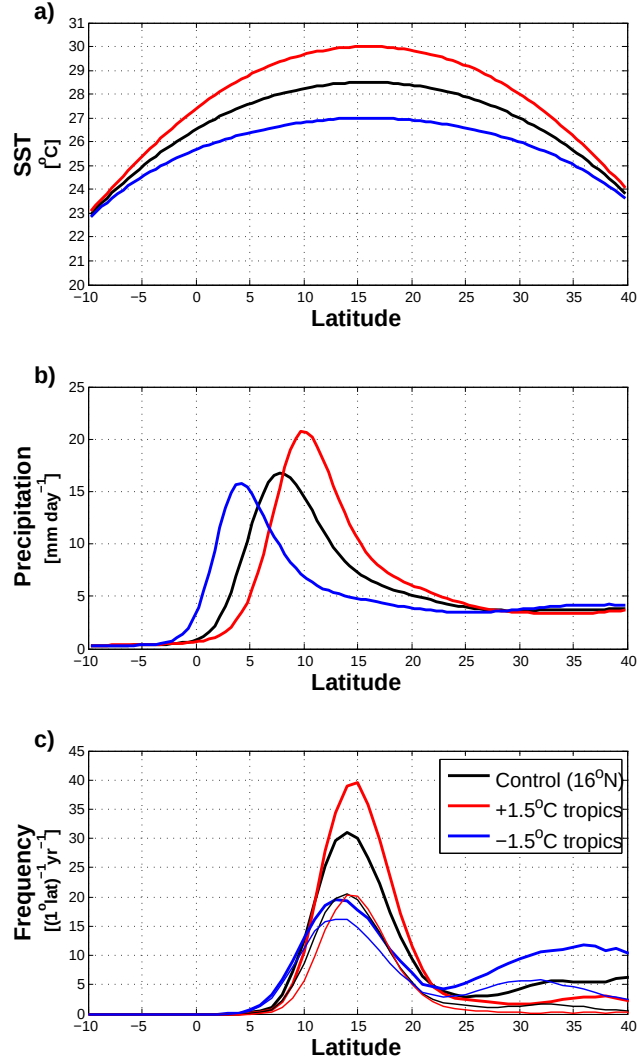


Figure 3.5: Time- and zonal-mean fields from the HiRAM aquaplanet fixed-SST simulations with a tropical SST perturbation: a) SST ($^{\circ}\text{C}$); b) precipitation rate (mm day^{-1}); and c) genesis frequency (TCs $(1^{\circ} \text{ lat})^{-1} \text{ yr}^{-1}$). The distribution of genesis frequency is shown for all tropical cyclones ($v_{\text{sfc}} > 15.2 \text{ m s}^{-1}$, thick lines) and for the subset that attain hurricane-strength wind speeds ($v_{\text{sfc}} > 29.5 \text{ m s}^{-1}$, thin lines).

3.5 Latitude of TC formation

The different simulations show subtle but distinct changes in the preferred region of tropical cyclogenesis. Recall that the latitude of tropical cyclogenesis here refers to the location of the first instance when a developing disturbance has $v_{\text{sfc}} > 15.2 \text{ m s}^{-1}$. Only those TCs that are equatorward of 25° N at their first detection are included

in the subsequent analyses. This appropriately avoids contamination of the results by those disturbances that have a sub-tropical origin (this smaller and well-separated mode is evident at higher latitudes in Fig. 3.2c, Fig. 3.4c and Fig. 3.5c).

In Figure 3.6 the mean latitude of TC genesis is plotted against the latitude of the maximum SST (the latitude θ_0 prescribed in the fixed-SST simulations, or the latitude diagnosed from the slab ocean simulations), and also the latitude of the maximum precipitation rate ($^{\circ}$ N). Each marker reflects a different 5-yr simulation of the HiRAM aquaplanet, with colored markers denoting fixed-SST experiments with maximum SST at 10° N (green), 13° N (blue), and 16° N (pink), recall Fig. 3.2a. Upward (downward) triangular markers denote the positive (negative) tropical perturbation experiments that were conducted at all three latitudes. Square markers show results from the accompanying slab ocean experiments.

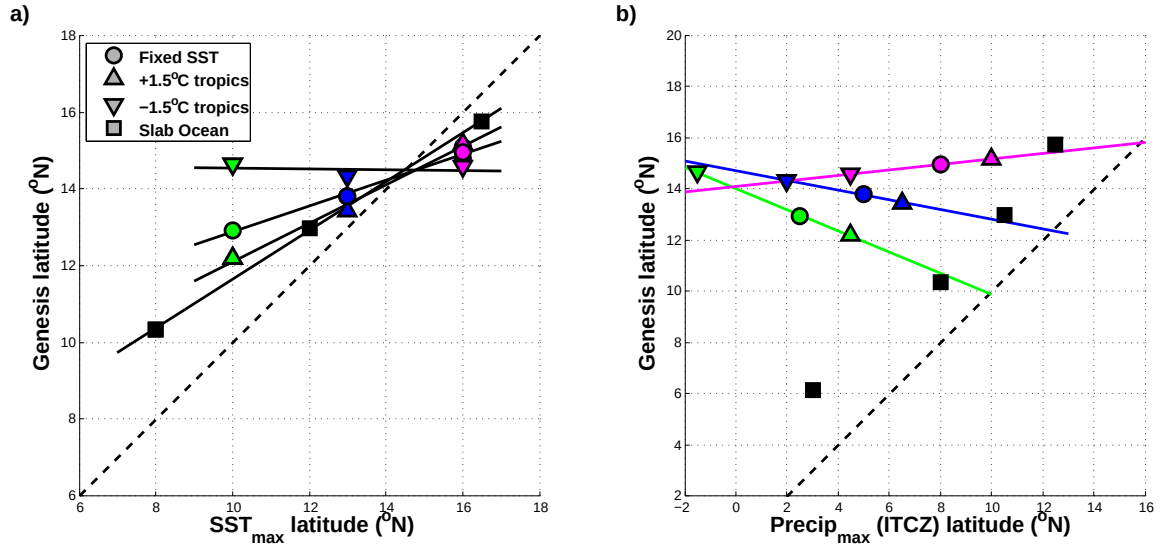


Figure 3.6: The mean latitude of tropical genesis plotted against the a) latitude of the maximum SST ($^{\circ}$ N), and b) latitude of the maximum precipitation rate as a measure of the location of the ITCZ ($^{\circ}$ N). Black solid lines indicate the linear fit through those simulations that have the same (or similar) tropical SST gradient. Each marker summarizes the TC statistics (genesis latitude $< 25^{\circ}$ N) from a 5-yr HiRAM aquaplanet simulation (see text for details).

Markers above and to the left of the 1:1 black dashed line in Figure 3.6a indicate preferred genesis poleward of the maximum SST, whilst markers below and to the right of the line have genesis primarily occurring equatorward of the maximum SST. The black solid lines indicate the linear fit through the different sets of simulations which share the same tropical SST gradient (or nearly the same SST gradient in the case of the slab ocean experiments where the SSTs are interactively determined).

The experiments which prescribe the flattest tropical SST gradients (those with cool tropical perturbations, SST_{T-}) show no sensitivity of the preferred genesis location to the latitude of the maximum SST; when the peak SST is shifted from 10°N to 16°N the resulting genesis remains centered at $\sim 14.5^\circ\text{N}$. However, in the sets of simulations with stronger SST gradients a sensitivity to the location of the SST maximum emerges as the preferred genesis region is observed to shift towards the latitude of the peak SST. The latitude of TC genesis tends toward the maximum SST as the SST gradient increases.

The observed linear relationship of the mean genesis latitude (θ_{GEN}) to the latitude of the maximum SST (θ_{SST}) within subsets of HiRAM aquaplanet simulations sharing similar SST gradients can be expressed simply as $\theta_{\text{GEN}} = \Gamma\theta_{\text{SST}} + c$. The sensitivity Γ appears to be related to the strength of the tropical SST gradient, such that $\Gamma \rightarrow 1$ as $|\nabla\text{SST}| \rightarrow \infty$, and $\Gamma \rightarrow 0$ as $|\nabla\text{SST}| \rightarrow |\nabla\text{SST}|_{\text{Threshold}}$ (i.e. zero once the SST gradient is sufficiently flat). Since the lines in Figure 3.6a are seen to intersect the approximate point $(14.5^\circ, 14.5^\circ)$, the preferred genesis latitude can be related to the latitude of the maximum SST by the simple expression, $\theta_{\text{GEN}} = 14.5^\circ + \Gamma(\theta_{\text{SST}} - 14.5^\circ)$.

Figure 3.6b shows the relationship between the latitude of TC genesis and the location of the ITCZ (θ_{ITCZ}), denoted here as the latitude of the maximum in time-mean precipitation rate within each of the simulations. Genesis is centered on the poleward flank of the ITCZ (above and to the left of the dashed line) in all experiments. In simulations where the prescribed latitude of the maximum SST is fixed (those with

the same colored markers), genesis occurs closer to the ITCZ in the simulations with a warm tropical perturbation (SST_{T+}) and further away from the ITCZ in simulations with a cool tropical perturbation (SST_{T-}). Furthermore, the linear fit lines through these subsets of experiments show that a systematic shift of θ_{GEN} accompanies the movement of θ_{ITCZ} .

For each of the colored lines in Figure 3.6b we observe that $\theta_{\text{GEN}} \rightarrow \theta_{\text{SST}}$ as $\theta_{\text{ITCZ}} \rightarrow \theta_{\text{SST}}$, and $\theta_{\text{GEN}} \rightarrow \sim 14.5^\circ$ as $\theta_{\text{ITCZ}} \rightarrow 0^\circ$, hence one can write the general linear expression,

$$\theta_{\text{GEN}} \approx 14.5^\circ + \frac{\theta_{\text{ITCZ}}}{\theta_{\text{SST}}} (\theta_{\text{SST}} - 14.5^\circ) , \quad (3.1)$$

relating the genesis latitude to the latitude of the maximum SST and ITCZ alone. Hence the aforementioned sensitivity of θ_{GEN} to shifts in θ_{SST} (the slopes of the black lines in Fig. 3.6a) can be approximated $\Gamma \sim \theta_{\text{ITCZ}}/\theta_{\text{SST}}$.

The results from both panels of Figure 3.6 confirm that the ITCZ shifts towards (away from) the maximum in SST as the tropical meridional SST gradient increases (decreases). Furthermore, within the HiRAM aquaplanet configuration, TC genesis remains centered around 14.5° N unless the latitude of the maximum SST is a sufficient distance away from 14.5° N *and* the tropical meridional SST gradient is strong enough to draw the ITCZ away from the equator toward the latitude of the maximum SST.

The simulations results may be interpreted as $\sim 14.5^\circ$ N being a *preferred* latitude for genesis in the HiRAM aquaplanet in the limit of a vanishing SST gradient. Planetary vorticity is the likely reason for the off-equatorial preferred genesis latitude. The bulk of TC genesis in the globally-uniform SST simulation of Shi and Bretherton (2014) appears to also occur near 15 degrees latitude. While this similarity is suggestive, the extent to which such a latitude is robust across different models or depends on the tracking scheme parameters is uncertain.

3.6 Discussion of TC activity

The accumulated cyclone energy (ACE; Bell et al., 2000; Camargo and Sobel, 2005) of an individual TC track is defined as the square of the estimated maximum sustained 10-m wind speed summed over all 6-hr periods from genesis ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$) through the end of the cyclone’s track,

$$\text{ACE}_i = \sum_{t=t_{\text{genesis}}}^{t_{\text{end of track}}} (\mathbf{v}_{\text{sfc}}^2)_t \quad [\text{m}^2\text{s}^{-2}]. \quad (3.2)$$

The total $\text{ACE} = \sum_{i=1}^N \text{ACE}_i$, summed over all analyzed TCs per year of model simulation will be used henceforth to broadly characterize the overall TC *activity* within each experiment. This measure of total TC activity depends on the TC genesis *frequency* (N TCs yr^{-1}), duration, and the mean ACE, a measure of the *average intensity* of TCs within a simulation,

$$\overline{\text{ACE}} = \frac{1}{N} \sum_{i=1}^N \left(\frac{\text{ACE}_i}{\Delta t_i} \right) \quad [\text{m}^2\text{s}^{-2} \text{ TC}^{-1} \text{ day}^{-1}]. \quad (3.3)$$

Note that $\overline{\text{ACE}}$ has here been computed after normalizing by the duration of the individual cyclone tracks ($\Delta t = t_{\text{end of track}} - t_{\text{genesis}}$) so as to present a metric that, when compared between experiments, focuses on changes in the maximum wind speed alone. As in the previous section, only TCs that form equatorward of 25°N are included in the analysis to ensure the focus remains on the tropical disturbances that are associated with the northern flank of the ITCZ.

3.6.1 Dependence on the SST and ITCZ configuration

The impact of the meridional SST profile and associated latitude of maximum precipitation rate (ITCZ) on the resulting TC statistics for each of the simulations are summarized in Figure 3.7. The upper panels (Fig. 3.7a,b,c) display the total TC activ-

ity ($\text{ACE}; \times 10^5 \text{ m}^2 \text{ s}^{-2} \text{ yr}^{-1}$), the middle panels (Fig. 3.7d,e,f) display the TC genesis frequency (yr^{-1}), and the lower panels (Fig. 3.7g,h,i) display the average intensity of TCs, normalized by track duration ($\overline{\text{ACE}}; \times 10^3 \text{ m}^2 \text{ s}^{-2} \text{ TC}^{-1} \text{ day}^{-1}$). The same convention for identifying different experiments has been adopted (as in Fig. 3.6). Also shown here are the results from the two globally-uniform SST perturbation simulations at 16° N , marked with a ‘+’ (SST_{G+}) and ‘×’ (SST_{G-}) symbol.

The TC statistics are plotted against the latitude of the maximum SST in the left panels (Fig. 3.7a,d,g). A clear relationship is observed in these experiments, **the total ACE increases as the latitude of the maximum SST moves poleward (Fig. 3.7a)**. In the fixed-SST control and slab ocean simulations this increase is realized in the frequency of TC formation (Fig. 3.7d) as well as in the average intensity of the cyclones (Fig. 3.7g). Note that for the same latitude of θ_{SST} , there is greater spread between the different experiments in the TC frequency and average intensity metrics when compared with the tighter fit seen in the overall product of these quantities. For example, while the fixed-SST experiments yield significantly fewer TCs than the slab ocean simulations, the $\overline{\text{ACE}}$ per fixed-SST TC (in the control experiments) is about twice that of the slab ocean TCs, resulting in a close relationship between θ_{SST} and total ACE across the different aquaplanet configurations. Similarly, in the fixed-SST simulations with a positive (negative) tropical SST perturbation, an increase (decrease) in the genesis frequency for a certain prescribed latitude of θ_{SST} , is largely offset by a decrease (increase) in the average intensity of those TCs such that the total ACE remains relatively unchanged. In contrast the average intensity of the TCs in the globally-uniform warmer (cooler) model simulations remains unchanged while the frequency decreases (increases). This results in an overall decrease (increase) of TC activity.

The TC statistics are plotted against the latitude of the maximum precipitation rate in the center panels (Fig. 3.7b,e,h). The prescribed SST profile plays a key role

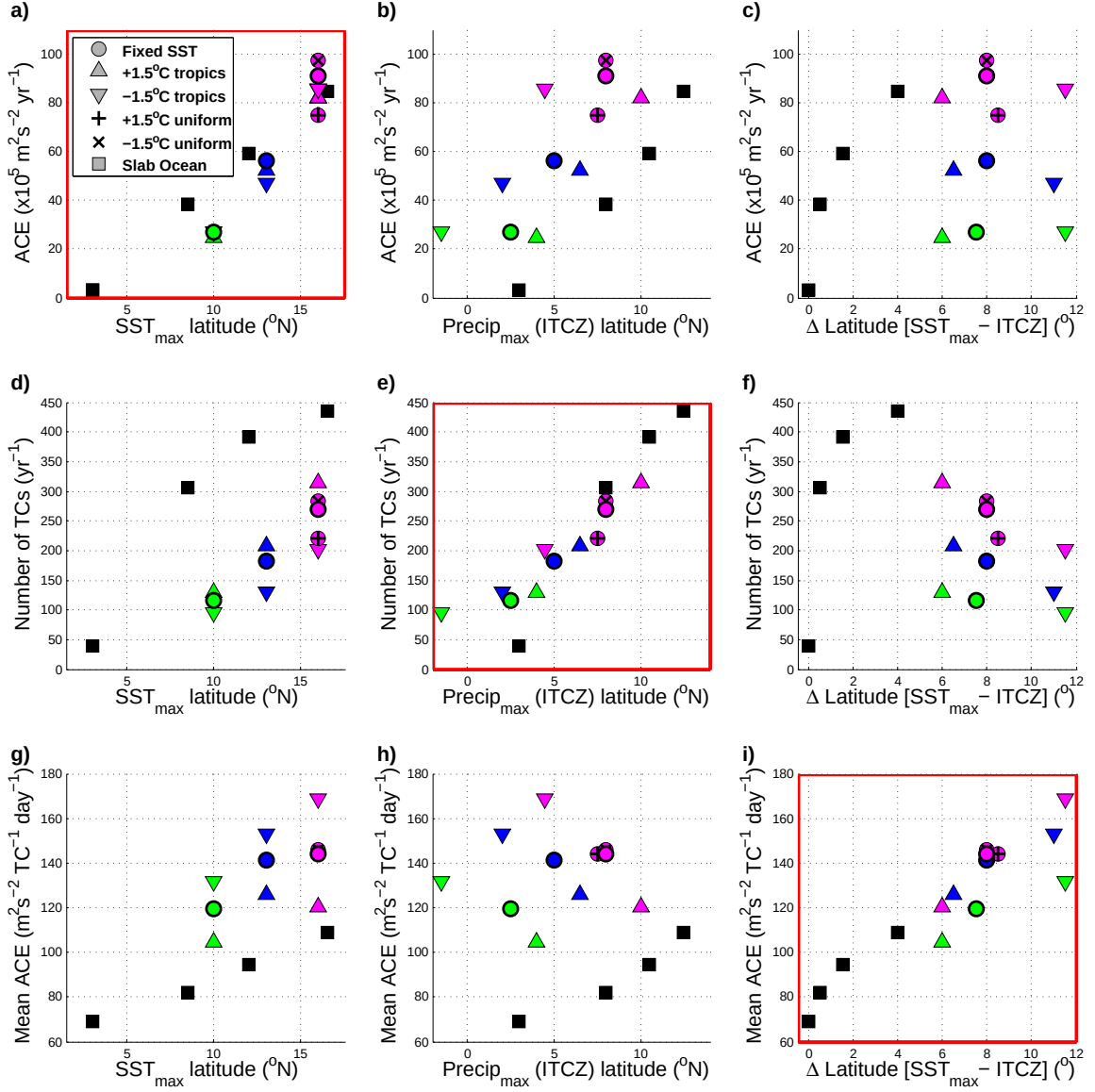


Figure 3.7: Tropical cyclone frequency and accumulated cyclone energy, plotted against the latitude of the maximum SST (left panels; a,d,g), the latitude of the maximum precipitation rate (center panels; b,e,h), and the latitudinal separation between those maxima (right panels; c,f,i). Upper panels (a,b,c) display the total ACE ($\times 10^5 \text{ m}^2 \text{ s}^{-2} \text{ yr}^{-1}$), middle panels (d,e,f) display the TC genesis frequency (yr^{-1}), and lower panels (g,h,i) display the mean ACE ($\text{m}^2 \text{ s}^{-2} \text{ TC}^{-1} \text{ day}^{-1}$). Each marker summarizes the TC statistics (genesis latitude $< 25^\circ \text{ N}$) from a 5-yr HiRAM aquaplanet simulation (see text for details).

in setting the location of the ITCZ in these aquaplanet simulations so it is unsurprising that the increasing TC activity, frequency, and average intensity observed with shifting the SSTs poleward (green $\rightarrow$ blue $\rightarrow$ pink markers) is also observed when

the statistics are plotted against the latitude of the ITCZ. However, as discussed earlier (Sec. 3.5), the *shape* of the meridional SST profile also influences the latitude of the ITCZ in these simulations, reflected here in the horizontal spread of the colored markers. Sharper (flatter) SST profiles move the ITCZ northward (southward) and toward (away from) the latitude of the maximum SST. While the total ACE in these experiments (Fig. 3.7b) appears to be relatively insensitive to changes in the meridional SST gradient alone, evidently **the frequency of TC genesis increases as the ITCZ latitude moves poleward (Fig. 3.7e)**. Correspondingly, for the same latitude θ_{SST} , a northward shift in the ITCZ (towards the SST maximum) is accompanied by a reduction in the average intensity of the simulated TCs (Fig. 3.7h).

This change in average intensity becomes clearer when the TC statistics are further plotted against the latitudinal separation between the location of the maximum SST and the ITCZ, $|\theta_{\text{SST}} - \theta_{\text{ITCZ}}|$, shown in the right panels (Fig. 3.7c,f,i). This abscissa provides a simple measure of the relative strength of the meridional SST gradient in driving the intertropical convergence toward the latitude of the maximum SST; the stronger the meridional SST gradient, the smaller the separation. As the latitudinal separation increases there is little sensitivity in the total TC activity (Fig. 3.7c) or frequency (Fig. 3.7f) metrics, however the average intensity is observed to increase robustly (Fig. 3.7i). Hence, in these aquaplanet experiments **the average intensity of TCs increase as the meridional temperature gradient flattens (Fig. 3.7i)**. This is consistent with the increase in the fraction of TCs that reach hurricane strength shown in Fig. 3.5c.

3.6.2 Dependence on environmental variables

Many previous investigations have sought to link the observed rate of TC genesis to the mean environmental ingredients known to be conducive for tropical cyclones. The quantities typically included in these tropical cyclongenesis indices include both

thermodynamic quantities, such as SST (absolute or relative to a tropical mean), potential intensity and relative humidity, and also kinematic quantities such as vertical wind shear and absolute vorticity. We do not seek here to rigorously test the various published indices (e.g., Emanuel and Nolan, 2004; Tippett et al., 2011) or propose a new index derived from these experiments. Rather, we give an overview of the relationship of select mean environmental quantities and the resulting frequency of TCs realized across this set of simulations.

To capture the mean value of a particular environmental field coincident at the genesis locations in these simulations we follow the convention introduced by Held and Zhao (2011) by defining the “genesis-weighted” quantity $\langle X \rangle_G$ here as

$$\langle X \rangle_G = \frac{\overline{XG}}{\overline{G}}, \quad (3.4)$$

where $X(\theta)$ is the zonal-mean quantity of interest and $G(\theta)$ is the histogram of genesis latitudes from the corresponding simulation; both $X(\theta)$ and $G(\theta)$ are recorded as monthly mean quantities. The overbar indicates the global mean (cosine-weighted latitude) of the quantities, further averaged (in time) over the whole simulation.

The frequency of TC formation in the HiRAM aquaplanet simulations is plotted against the genesis-weighted absolute sea surface temperature $\langle SST \rangle_G$ in Figure 3.8a. Note the large vertical spread in observed frequency for a given value of average underlying SST (e.g., $\sim 100\text{--}400$ TCs yr^{-1} at 28°C). These results confirm that absolute SST alone (averaged over the genesis region) is not the dominant factor governing the number of simulated TCs in these experiments. Qualitatively similar results are found if tropical- or global-mean SST is plotted as the abscissa. While Held and Zhao (2011) found a TC frequency sensitivity of $\sim -5.5\% \text{ K}^{-1}$ global mean surface temperature in HiRAM simulations, Merlis et al. (2013) showed that the TC frequency sensitivity to warming could be appreciably masked or enhanced by a coincident change in the

mean latitude of the ITCZ. Across our experiments there are significant changes in the ITCZ time-mean latitude (along with its width and strength), and hence these movements likely contribute greatly to the observed changes in TC genesis. The exception is in the case of the globally uniform perturbation simulations ($SST_{G\pm}$) which can be appropriately compared with the 16°N control (pink circle markers) as there is little change to the latitude of the ITCZ between those simulations. This comparison confirms a decreasing tendency of genesis with a uniform increase in SST alone.

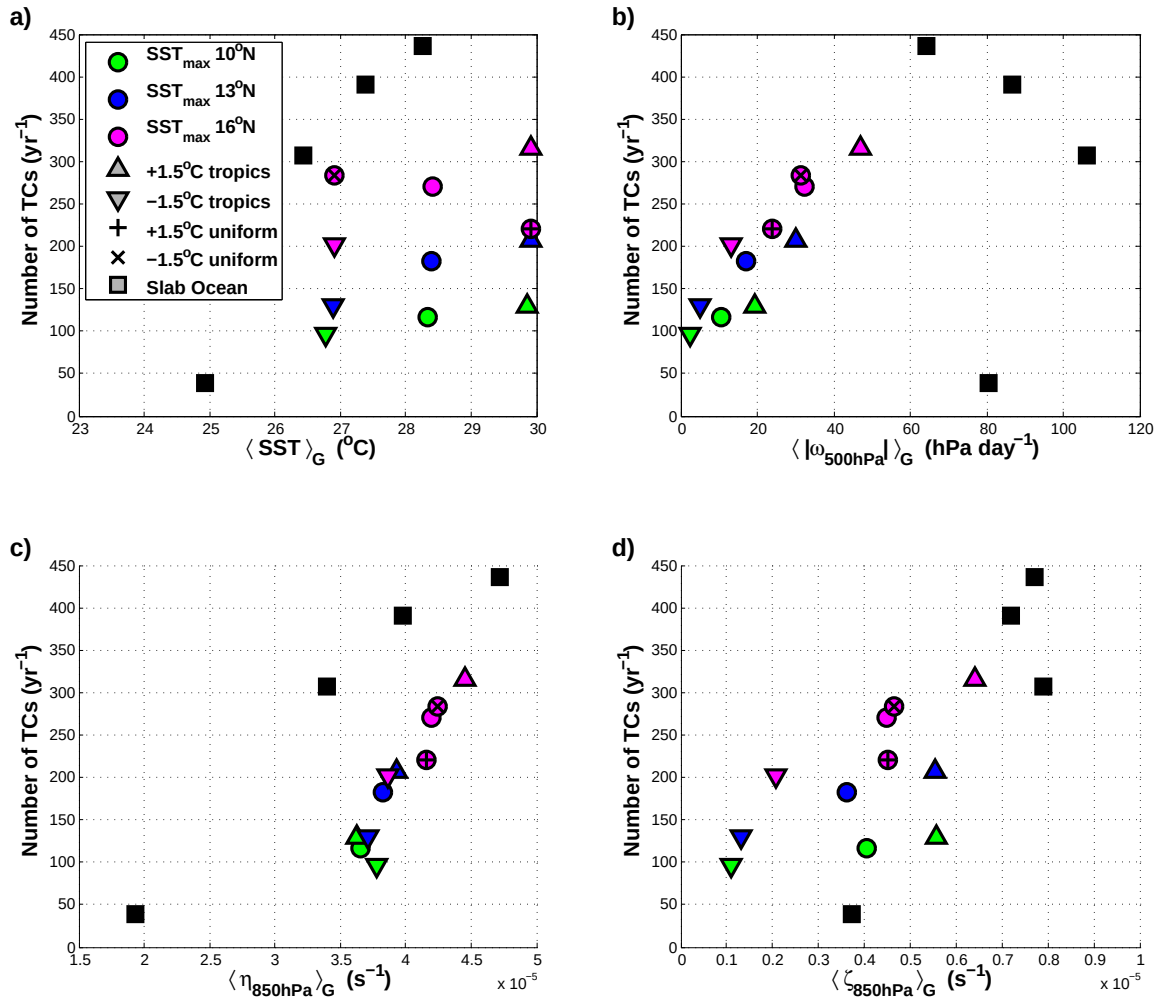


Figure 3.8: Tropical cyclone frequency (yr⁻¹) plotted against several genesis-weighted environmental variables: a) SST (°C); b) magnitude of the 500-hPa vertical velocity (hPa day⁻¹); c) 850-hPa absolute vorticity ($\eta = f + \zeta$, s⁻¹); and d) 850-hPa relative vorticity (ζ , s⁻¹). The $\langle \rangle_G$ notation indicates the spatial and time mean quantities over the region of TC genesis.

The strength of the mean upward mass flux has been previously linked to changes in TC genesis frequency (e.g., Held and Zhao, 2011; Sugi et al., 2012; Sugi and Yoshimura, 2012), and here we show the genesis-weighted magnitude of the 500-hPa pressure velocity $\langle |\omega_{500}| \rangle_G$ (Fig. 3.8b). TC genesis frequency is observed to increase with $|\omega_{500}|$ in fixed-SST simulations, but this relationship is not apparent in the accompanying slab ocean experiments. While in Held and Zhao (2011) a clear relationship was found between mean mass flux and TC frequency when examining one environmental field at a time, Camargo et al. (2014) more recently used vertical velocity as one of the predictors in a set of more comprehensive genesis indices for the simulations of the HiRAM model in present and future climate, and found that the inclusion of the vertical velocity did not explain the changes in TC frequency in the future climates.

One can question whether the strength of the vertical motion in the genesis region is controlled by the TC activity rather than the converse. Zhao et al. (2012) have discussed how TC genesis in a version of HiRAM with realistic SSTs can be reduced dramatically by smoothing the horizontal divergence field, without modifying the distribution and mean vertical motion significantly. The implication is that the latter is hardly affected by the amount of TC activity. It will be of interest to perform similar parameter sensitivity studies of this and other types in the future to explore these issues more fully.

Along with the strength of the convergence, the latitude of the ITCZ is also an important influence on the number of developing disturbances (Fig. 3.7e). A standard parameter typically included in various TC genesis indices is a measure of the absolute vorticity $\eta = f + \zeta$, the sum of the latitude-dependent planetary vorticity $f = 2\Omega \sin \theta$ and the relative vorticity ζ of the environmental flow. The genesis-weighted 850-hPa absolute vorticity $\langle \eta_{850} = \zeta_{850} + f \rangle_G$ for the various HiRAM simulations is shown in Figure 3.8c.

The different HiRAM simulations show a compact and correlated relationship of increasing TC genesis frequency with increasing absolute vorticity, although the slab ocean and fixed-SST experiments do not align on the same curve. The relatively higher frequency of TC genesis in the slab ocean simulations for similar values of $\langle \eta_{850} \rangle_G$ in the fixed-SST simulations (Fig. 3.8c) may help to partially explain the discrepancy in $\langle |\omega_{500}| \rangle_G$ (Fig. 3.8b), where higher values of upward mass flux in the slab ocean simulations do not yield an increase in TC genesis frequency consistent with the relationship suggested by the fixed-SST experiments. An index that combined proxies for both the strength and latitude of the ITCZ would thus likely yield a tighter relationship. The results shown here are consistent with the “threshold”-type behavior discussed in McGauley and Nolan (2011), where TCs are unable to form without a sufficient value of environmental absolute vorticity.

The changes in $\langle \eta_{850} \rangle_G$ between different experiments largely reflects the changes to the mean latitude of TC genesis (recall Fig. 3.6) and the associated ambient planetary vorticity of the environment. A smaller part of the change in $\langle \eta_{850} \rangle_G$ comes from changes in the relative component of vorticity $\langle \zeta_{850} \rangle_G$ shown in Figure 3.8d, which is associated with the structure of the time-mean zonal flow ($-dU/dy$) on the poleward flank of the ITCZ. Here the relative vorticity appears to be more sensitive to changes in the shape of the SST profile than to shifts in the latitude of the maximum SST. Changes in the zonal-mean flow may be important in setting the stability of the ITCZ and the propensity for breakdown and generation of TC-like disturbances (e.g., Nieto Ferreira and Schubert, 1997).

Motivated by our earlier finding (Sec. 3.5) that suggested a relatively flat-SST environment (with the ITCZ remaining close the equator) would prefer to form TCs centered about the latitude of 14.5°N , we consider finally here the environment of this region across all experiments without concerning ourselves with knowing where genesis *actually* occurs. Figure 3.9 shows the potential utility of focusing solely (for

simplicity) on the environment centered over 14.5°N . Here the time-mean upward vertical motion $-\omega_{500}$ has been spatially averaged over a latitude band centered at 14.5°N . A Gaussian kernel of width 10° latitude has been used for the spatial averaging, but in fact the result is very similar to that shown if one only analyzes the time-mean $-\omega_{500}$ at the gridbox nearest to 14.5°N (without any latitudinal averaging). The number of TCs that form across the various configurations of the HiRAM aquaplanet we have analyzed show a striking relationship with the upward mass flux over the 14.5°N latitude region. Moreover, the relationship is shown to tighten further if an adjustment is made to the TC frequency to account for changes in the mean tropical SST between the experiments (e.g., $-5\% \text{K}^{-1}$, not shown here).

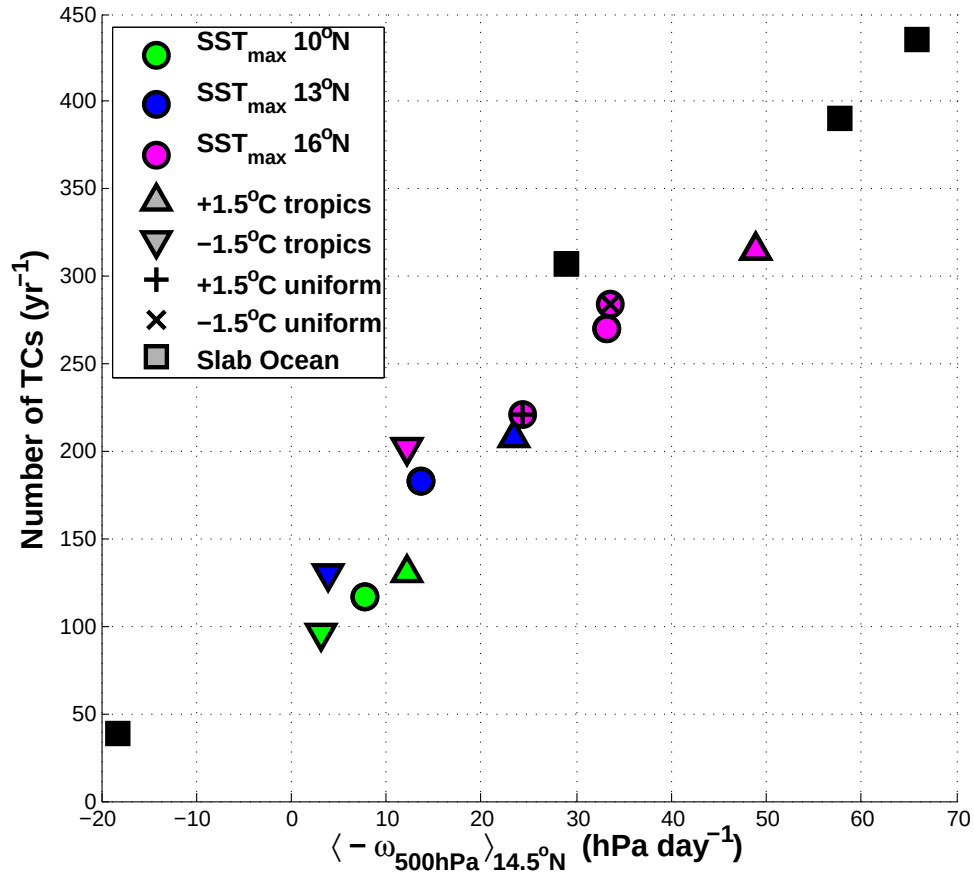


Figure 3.9: Tropical cyclone frequency (yr^{-1}) plotted against the time-mean vertical velocity (hPa day^{-1}), spatially averaged over a region centered at 14.5°N , where TC genesis occurs in the limit of weak SST gradient.

3.7 Summary

In the zonally symmetric, interhemispherically asymmetric climates of the HiRAM aquaplanet presented in this chapter, tropical cyclones form on the poleward flank of the off-equatorial ITCZ, subsequently strengthening and decaying in their intensity as they propagate to higher latitudes. When analyzing the TC activity in these simulations it is a challenge to separate any influences of the prescribed SST profile from that of the latitude of the ITCZ; the ITCZ latitude is of course related to the SST profile in some nontrivial way. However, by fixing some aspect of the SST profiles between experiments (e.g., the magnitude of the maximum SST and ∇ SST) while adjusting some other characteristic (e.g., the latitude of the maximum SST), a set of aquaplanet climatologies were formed, each with a small perturbation to the SST profile and associated changes in the latitude of the ITCZ.

The resulting variation in the genesis latitude, frequency and intensity have been investigated across this range of SST-ITCZ environments in order to identify consistent and robust changes in the statistics of simulated TCs. We find in these simulations that when the tropical meridional SST profile is sufficiently flat that the TCs typically form near $\sim 14.5^\circ$ N. However, as the tropical meridional SST gradient increases, both the ITCZ and genesis region shift towards the latitude of the maximum SST.

The total TC activity (defined here by the Accumulated Cyclone Energy) increases as the latitude of the maximum SST moves poleward. This occurs due to increases in both the genesis frequency and the average intensity of the TCs that form as the maximum SST location shifts poleward. The simulated increase in the genesis frequency may be attributed to the concurrent and related poleward shift in the latitude of the ITCZ. The experiments which move the ITCZ further poleward while holding the latitude of the maximum SST fixed (through changes in the magnitude of ∇ SST) result in a higher genesis frequency. These experiments also reveal that

the TCs which form in environments where the ITCZ is further poleward and closer to the latitude of the maximum SST are on average less intense, and hence the total activity is relatively insensitive to the changes in the latitude of the ITCZ alone. We recognize that storm intensities in this model are a function of resolution, so one must assume that results on statistics such as the accumulated cyclone energy will change quantitatively at higher resolution. Whether the qualitative results described here will change at higher resolution remains to be seen.

While the strength of mid-level ascent ($|\omega_{500}|$) has been shown here and elsewhere to be an important mediator of TC frequency, the upward mass flux over the genesis region alone is not determinative of TC frequency, as notably demonstrated in the slab ocean experiments. Changes in the latitude of the genesis region, and thus the absolute vorticity available to developing TCs, also influence the genesis frequency. Although beyond the scope of this investigation, convectively-coupled waves that propagate throughout the tropical region, along with instabilities that develop along the ITCZ itself are likely important in the formation of the precursor disturbances, from which a certain fraction develop into self-sustaining tropical cyclones. The extent to which the large-scale influences the fraction of disturbances that develop into tropical cyclones still remains an open question.

Chapter 4

TC intensity evolution

4.1 Motivation

In the previous chapter the focus was on exploring the frequency of TC genesis in the HiRAM aquaplanet simulations, with less regard for analyzing the subsequent evolution of the disturbances once they develop into TCs. However, as was evidenced in the discussion of the total TC activity and average intensity (Sec. 3.6), it is difficult to completely separate the climatological factors influencing the propensity of forming TCs (genesis frequency), from those factors governing (or at least influencing) the subsequent growth and/or decay of the individual TCs. For example, more favorable (or unfavorable) environmental conditions might prolong (or shorten) the duration of TCs, impacting the statistics of these events and perhaps even the number of TCs captured by the definition in the first place.

Certainly one must consider the large-scale environment that the individual TCs traverse throughout their lifetime in order to better understand the differences in lifetime maximum intensity or total TC activity (measured by ACE, for example). Moreover, the HiRAM aquaplanet provides a suitable (and advantageous) arena for investigating the evolution of TCs, because from the outset one avoids the complicat-

ing influence of surface inhomogeneities that could otherwise play a dominant role in affecting the growth or decay of TCs. In contrast, observational datasets or GCMs applying more realistic (complicated) boundary conditions will enable a clearer analysis of basin-to-basin variability and landfall statistics, of which surface inhomogeneities obviously play a critical role. Rather, with the zonally-symmetric aquaplanet our focus can be on identifying features of the large-scale climate that are influencing the evolution of TCs, without the muddying effect (statistically, at least) of accounting for the differing surface conditions. The zonally-symmetric conditions allow the evolution of the individual TCs to be compared in a simple latitude-time framework.

This chapter is also motivated in part by a recent study by Kossin et al. (2014) that explored the latitude at which observed TCs acquired their lifetime maximum intensity. This particular metric ought to be useful in detecting the influence of different climate regimes on TC behavior since one could reasonably expect the observational record to be more reliable in capturing a tropical cyclone’s time and location of maximum intensity than the time and location of genesis or lysis, for example. Our zonally-symmetric framework provides an ideal opportunity to investigate the latitude of maximum intensity across our suite of perturbation experiments, in order to say something about the large-scale environment’s influence on this aspect of TC activity.

4.2 Results

The mean time evolution of the intensity of tropical cyclones in the simulations can be visualized by a sequence of markers that are equally spaced in time, plotted on a graph with the mean wind speed as the abscissa and mean latitude as the ordinate (see Fig. 4.1, for example). The sequence of markers begin in the lower left portion of the graph, and reflects the mean wind speed and mean latitude of all TCs (within

a particular simulation) at the moment of genesis (first detection). Subsequent small markers are plotted every 6 hr (larger solid markers are plotted every 1 day), showing the mean wind speed and latitude of the TCs throughout the first 14 days of their evolution.

Although there is a large spread in the distribution of actual latitudes and wind speeds that the individual TCs have throughout their lifetime (not shown), an indication of the confidence in the position of the mean of the distribution is given by the length of the vertical and horizontal line segments. The line segments indicate the 5-95% confidence interval for the estimates of the mean latitude and mean intensity, determined from bootstrap sampling ($N = 1000$) of the statistic.

When exploring the evolution of TC intensity it will also be of interest to characterize the actual average maximum intensity of TCs within a simulation, $\left(\frac{1}{N} \sum_{i=1}^N \max(\text{TC}_i)\right)$, and the associated mean latitude at which this maximum intensity is observed to occur. Note that this measure (average maximum intensity) is subtly different from the maximum intensity of the average TC trace that is deduced from the aforementioned temporal-evolution plots.

For example, consider the two following sequences of numbers, $\text{TC}_1 = \{25, 35, 25, 20\}$ and $\text{TC}_2 = \{15, 15, 35, 30\}$, designed to represent the simple evolution of wind speed from two different (hypothetical) tropical cyclones at days 1 through 4. The maximum intensity of TC_1 is 35 and occurs on the second day; for TC_2 the maximum is also 35 but it occurs on the third day. The actual average maximum intensity of these two TCs is $\frac{1}{2} \sum_{i=1}^2 \max(\text{TC}_i) = 35$. However, when considering the mean temporal evolution of these two example storms, $\text{TC}_{av} = (\text{TC}_1 + \text{TC}_2) / 2 = \{20, 25, 30, 25\}$, we note that the *average* TC reaches a maximum intensity of 30 at day 3. This important distinction must be kept in mind as the following results are assessed.

4.2.1 Poleward shift in maximum SST

Evolution of TC intensity

To illustrate the impact of a poleward shift in the latitude of the prescribed SST maximum, the mean time evolution of all TCs in the three control simulations is shown in Figure 4.1. The mean genesis latitude ($\theta_{\text{GEN}} = \sim \{13^\circ \text{ N}, 14^\circ \text{ N}, 15^\circ \text{ N}\}$) shifts poleward as the latitude of the maximum SST ($\theta_{\text{SST}} = \theta_0 = \{10^\circ \text{ N}, 13^\circ \text{ N}, 16^\circ \text{ N}\}$) is shifted poleward in the control experiments.

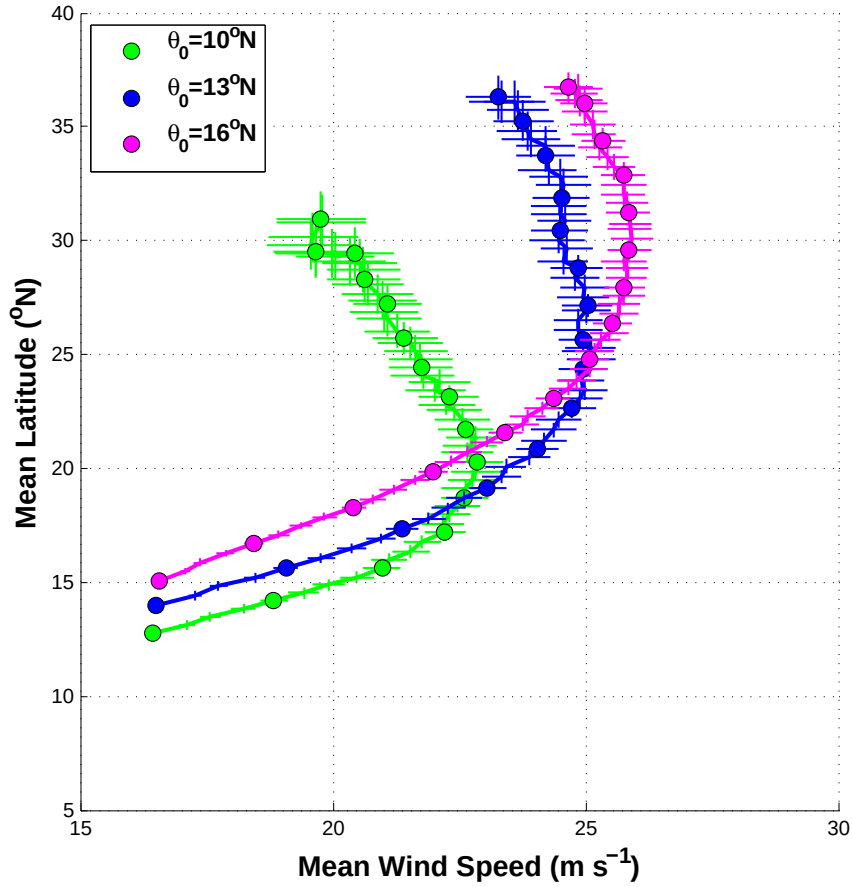


Figure 4.1: The time evolution of TCs in the HiRAM aquaplanet fixed-SST control simulations. Starting from genesis and plotting subsequent 6-hr values for a 14-day period, each cross shows the 5-95% bootstrap confidence interval for the estimates of the mean latitude ($^\circ \text{ N}$) and mean intensity (surface wind speed, m s^{-1}), as a function of track duration. Markers of the mean latitude and intensity are plotted every 1 day of track duration.

The first point to note from Figure 4.1 is that the simulated TCs tend to reach their maximum intensity later in their lifetime (subsequent to genesis) as the latitude of maximum SST is shifted polewards. For the first 2-3 days following genesis the average TCs move poleward ($\sim +1.5^\circ \text{ day}^{-1}$) and intensify ($\sim +2.5 \text{ m s}^{-1} \text{ day}^{-1}$) at a similar rate across the three different control simulations. Figure 4.2 shows the mean time evolution of frequency (a), latitude (b) and intensity (c) of the TCs following genesis (left panels).

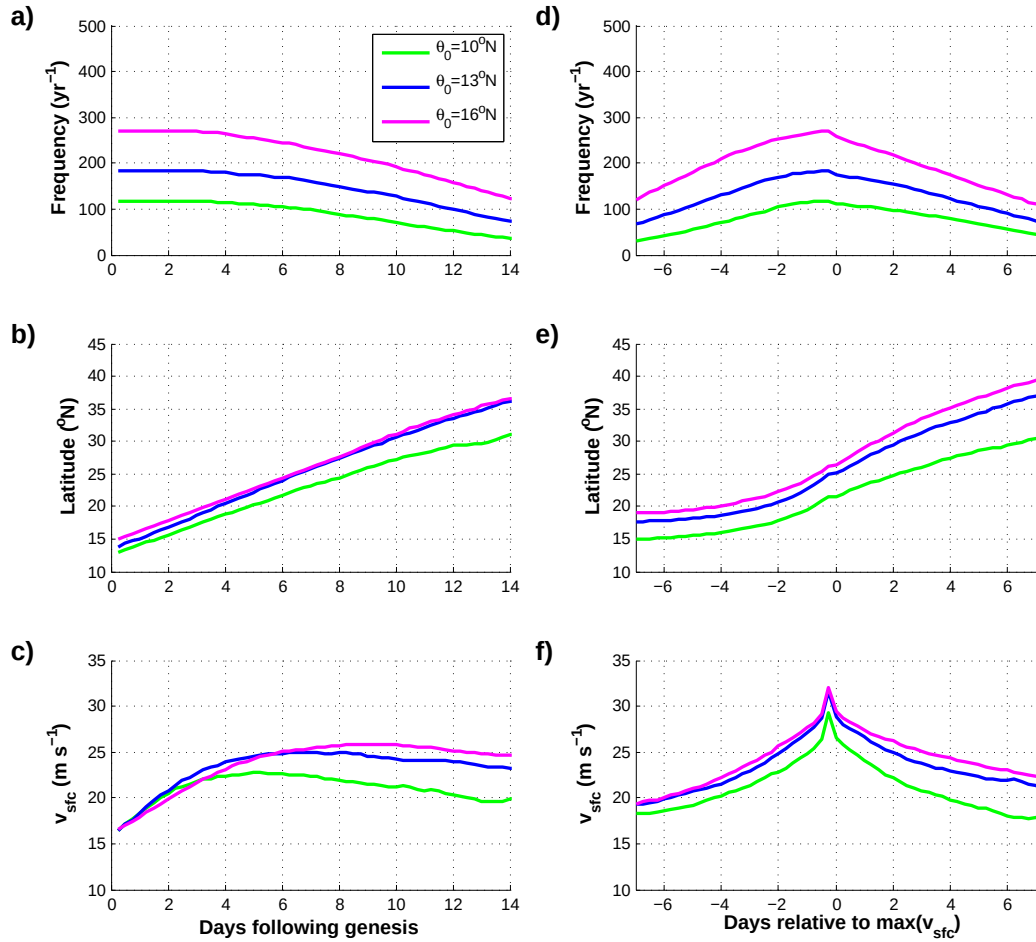


Figure 4.2: The time evolution of TCs in the HiRAM aquaplanet fixed-SST control simulations, with reference to the time of first detection [a,b,c], and to the time of the lifetime maximum intensity [d,e,f]. Plotted is the frequency (yr^{-1}) [a,d]; the mean latitude ($^\circ \text{N}$) [b,e]; and the mean intensity (m s^{-1}) [c,f] of the tracked disturbances.

The differences in the genesis frequency of simulated TCs have been previously discussed (Sec. 3.2), and Figure 4.2a suggests that the proportionate difference in frequency between the control experiments does not change remarkably over time. The average poleward translation rate (Fig. 4.2b) also remains relatively constant over the 14 days, slowing slightly in the $\theta_{\text{SST}} = 10^\circ \text{N}$ simulation. However, the average evolution of storm intensity differs significantly after the first 2-3 days (Fig. 4.1, Fig. 4.2c).

The average evolution of simulated TCs can also be plotted with reference to the day of each individual TC track's lifetime maximum intensity, as shown in Figure 4.2 (right panels). The plot displaying the evolution of TC latitude (Fig. 4.2e) shows that the time when the maximum intensity is reached (day = 0) is coincident with a maximum in the poleward latitudinal translation of TCs. For the several days preceding the maximum in TC intensity the convex-up curves indicate a poleward acceleration of TCs (on average); a slight deceleration is then seen in the days following the maximum intensity.

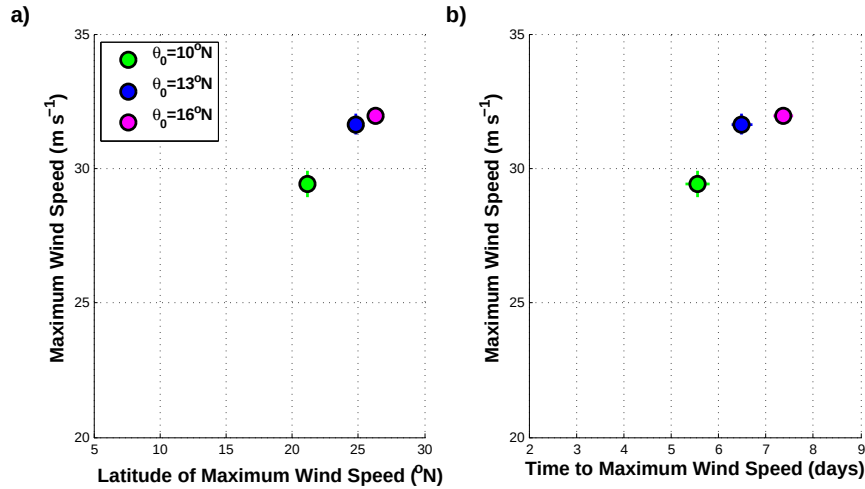


Figure 4.3: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet fixed-SST control simulations, plotted against the a) mean latitude corresponding to where the TCs achieve their maximum wind speed ($^\circ \text{N}$); and b) the time interval between genesis and when the TCs achieve their maximum wind speed (days). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

The actual average maximum intensity $\left(\frac{1}{N} \sum_{i=1}^N \max(\text{TC}_i)\right)$ within each of the control simulations is reflected by the values in Figure 4.2f at day = 0. These values of average lifetime maximum intensity are plotted in Figure 4.3 against the associated mean latitude of maximum intensity (a), and the mean time interval between genesis and the point at which the TCs achieve their maximum intensity (b).

Figure 4.3 shows that TCs in the $\theta_{\text{SST}} = 10^\circ \text{N}$ control simulation strengthen (on average) for ~ 5.5 days, acquiring a mean maximum wind speed of $\sim 29 \text{m s}^{-1}$ at a corresponding mean latitude of $\sim 21.5^\circ \text{N}$. As the latitude of the prescribed SST maximum shifts polewards, the TCs experience a longer intensification period, acquiring a higher mean lifetime maximum intensity at a higher mean latitude. In the northern-most control simulation ($\theta_{\text{SST}} = 16^\circ \text{N}$) TCs strengthen for ~ 7.5 days, acquiring a mean maximum wind speed of $\sim 32 \text{m s}^{-1}$ at a mean latitude of $\sim 27^\circ \text{N}$.

Evolution of the TC environment

The relationship of the large-scale environment to the evolution of TCs within the simulations is considered next. From the latitude recorded at every point along a simulated TC's track (6-hr output interval), the Lagrangian evolution of each TC's climatological environment can be retrieved from the time- and zonal-mean fields, linearly interpolated from the nearest model output latitudes. Sampling from the climatological fields avoids any influence from the disturbance itself.

The potential intensity (PI) of a tropical cyclone (TC) refers to its theoretical upper limit of intensity, given specified sea surface temperature and local profiles of temperature and humidity (Emanuel, 1986). Because PI can be computed readily from SST and a local thermodynamic profile alone, it is readily diagnosed from global climate models and is often used as a proxy for assessing how actual TC intensities may respond to climate change. PI is computed here following the method of Bister and Emanuel (2002); script at <ftp://texmex.mit.edu/pub/emanuel/TCMAX/>.

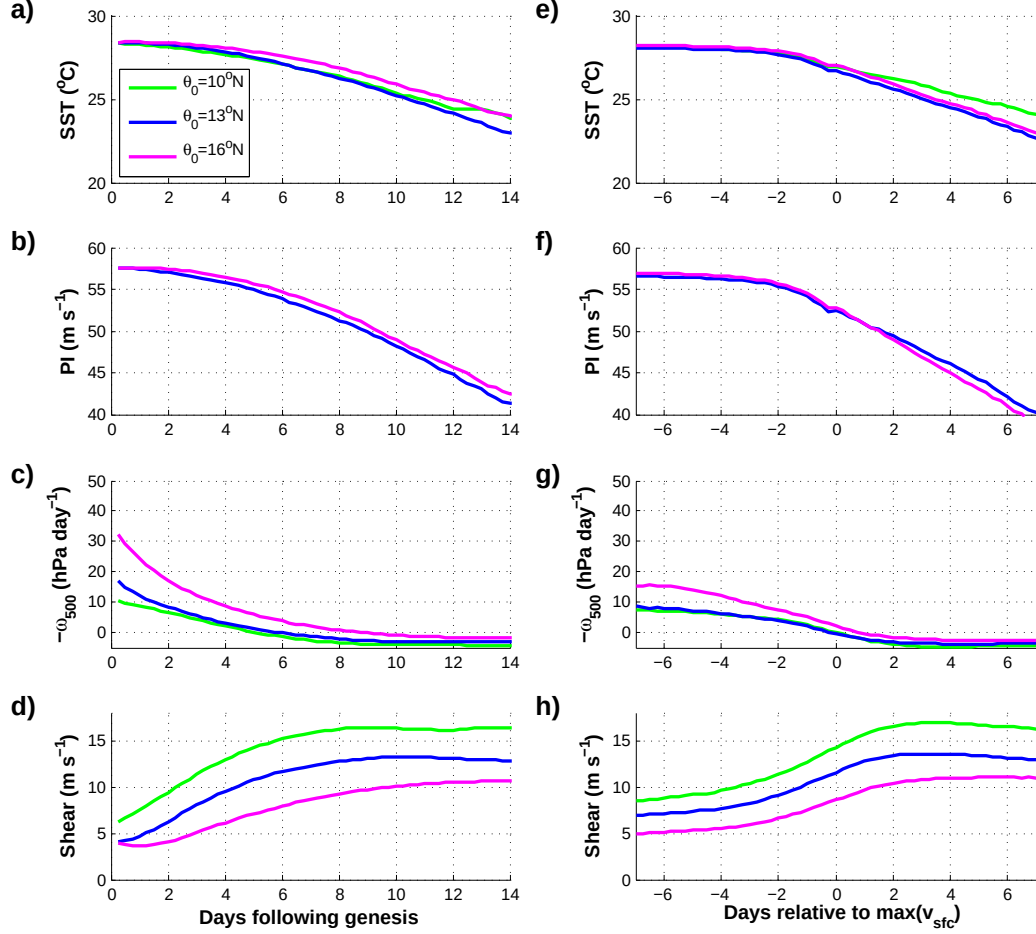


Figure 4.4: The time evolution of the the mean environmental conditions that the simulated TCs traverse throughout their lifetime, with reference to the time of first detection [a,b,c,d], and to the time of their lifetime maximum intensity [e,f,g,h]. Plotted is the SST ($^{\circ}\text{C}$) [a,e]; the Potential Intensity (m s^{-1}) [b,f]; the 500-hPa vertical pressure velocity (hPa day^{-1}) [c,g]; and the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}) [d,h].

Results are shown in Figure 4.4 for the evolution of sea surface temperature, Potential Intensity, vertical velocity and vertical wind shear fields. As expected from the prescribed boundary condition, the underlying SST (Fig. 4.4a) that the TCs *experience* throughout their lifetime reduces as the TCs migrate to higher latitudes (recall Fig. 4.2b). The evolution of PI (Fig. 4.4b, noting the $\theta_0 = 10^{\circ}\text{N}$ curve has been omitted) largely reflects the underlying SST changes, and thus reduces over the lifetime of the TCs.

The mean vertical velocity (Fig. 4.4c) of the TC environment is strongest at genesis when the disturbances initially form on the poleward flank of the ITCZ. As the TCs develop and move away from the ITCZ the mean vertical velocity of the large-scale environment reduces, shifting instead to a regime of gentle subsidence in the subtropics. The evolution of the magnitude of the 250-850 hPa vertical shear of the horizontal wind is shown in Figure 4.4d. TCs form in a region of relative low wind shear near the ITCZ, but experience progressively greater shear as they move to higher latitudes.

Figure 4.5 shows the mean value of SST, PI, vertical velocity and wind shear in the large-scale environment coincident with the TCs achieving their lifetime maximum intensity. The rate of decreasing PI increases over the two days leading up to the maximum intensity (Fig. 4.4f) in all experiments, associated with the mean increase in poleward motion (Fig. 4.2e). When comparing the different control experiments (as the latitude of the prescribed SST maximum is shifted polewards) there is no noticeable difference in the average evolution of the SSTs underlying the simulated TCs (Fig. 4.4e) or related PI (Fig. 4.4f) prior to the maximum in TC intensity being achieved. The day = 0 values of SST (Fig. 4.5a) and PI (Fig. 4.5b) are virtually the same, and thus the SST or PI cannot adequately explain the difference in the mean lifetime maximum wind speed of TCs in the different control simulations.

The mean upward vertical motion experienced throughout the period of TC intensification (Fig. 4.4g) reduces and changes to descent (on average) around the same time that TCs achieve their maximum intensity (Fig. 4.5c). The evolution of environmental vertical wind shear (Fig. 4.4h) differs substantially for TCs within the different control runs. From genesis through to the point of maximum intensity (Fig. 4.5d), the TCs in the $\theta_0 = 10^\circ \text{N}$ control simulation experience (on average) almost double the magnitude of large-scale vertical wind shear of those TCs in the $\theta_0 = 16^\circ \text{N}$ simulation.

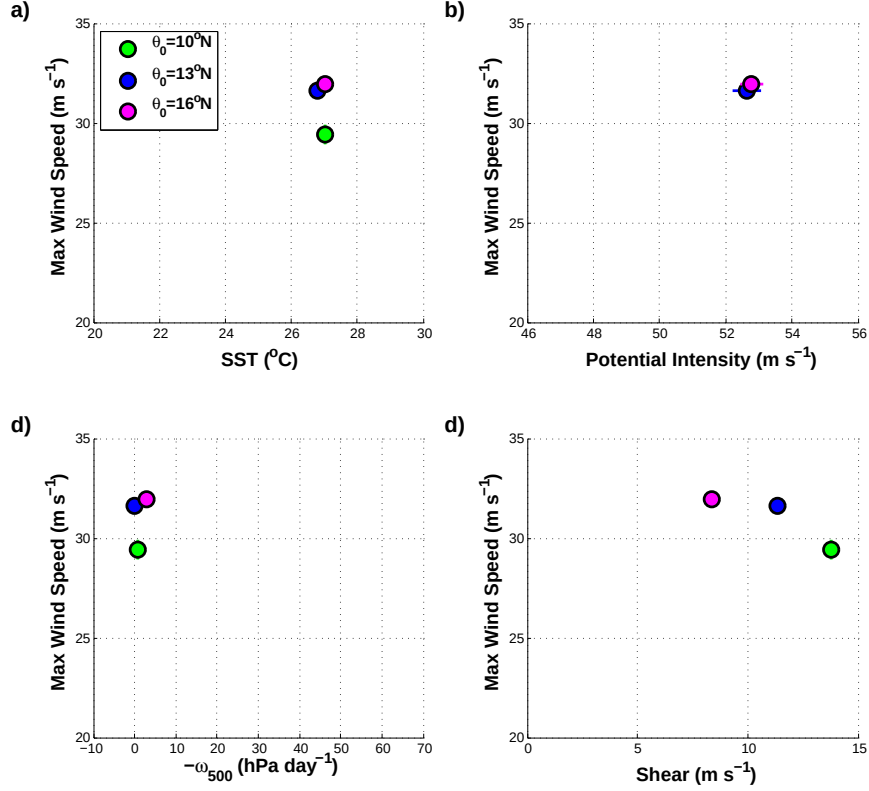


Figure 4.5: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet fixed-SST control simulations, plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity. Plotted is the a) SST (°C); b) Potential Intensity (m s⁻¹); c) 500-hPa vertical pressure velocity (hPa day⁻¹); and d) the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s⁻¹). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

Summary

As the latitude of the prescribed SST maximum in the HiRAM aquaplanet fixed-SST control simulations shifts poleward:

- The mean maximum intensity of simulated TCs increases.
- TCs reach their maximum intensity at higher latitudes.
- TCs have a longer period of intensification, reaching maximum intensity later in their lifetime.

- TCs experience an environment where the underlying SSTs and PI are very similar, the mean vertical velocity is slightly stronger, and the vertical wind shear is considerably weaker.

4.2.2 Global perturbation experiments

Evolution of TC intensity

To investigate the impact of a globally uniform SST perturbation, the mean time evolution of all TCs in the $\theta_0 = 16^\circ$ N control simulation, along with the two simulations corresponding to a 1.5° C uniform warming and 1.5° C uniform cooling, are shown in Figure 4.6. Similarly, Figure 4.7 shows the evolution of frequency (a), latitude (b), and intensity (c) of the TCs with reference to the genesis time (left panels).

While there are significant differences in the number of simulated TCs when a globally uniform warming or cooling is applied (recall Sec. 3.3), these frequency differences do not appear to change (proportionally) as a function of TC duration, whether plotted with respect to genesis (Fig. 4.7a), or in reference to the day of lifetime maximum intensity (Fig. 4.7d). The evolution of mean latitude (Fig. 4.7b,e) and intensity (Fig. 4.7c,f) are almost identical in the simulations with a globally uniform SST perturbation.

Figure 4.8 displays the mean maximum intensity plotted firstly against the mean latitude where the TCs acquire their maximum intensity (a), and then against the mean time interval between genesis and the point of maximum intensity (b). The mean values for the different experiments ($\pm 1.5^\circ$ C) are statistically indistinguishable. In all three simulations, subsequent to genesis, simulated TCs strengthen (on average) for ~ 7.5 days, acquiring a mean maximum wind speed of $\sim 32 \text{ m s}^{-1}$ at a mean latitude of $\sim 27^\circ$ N.

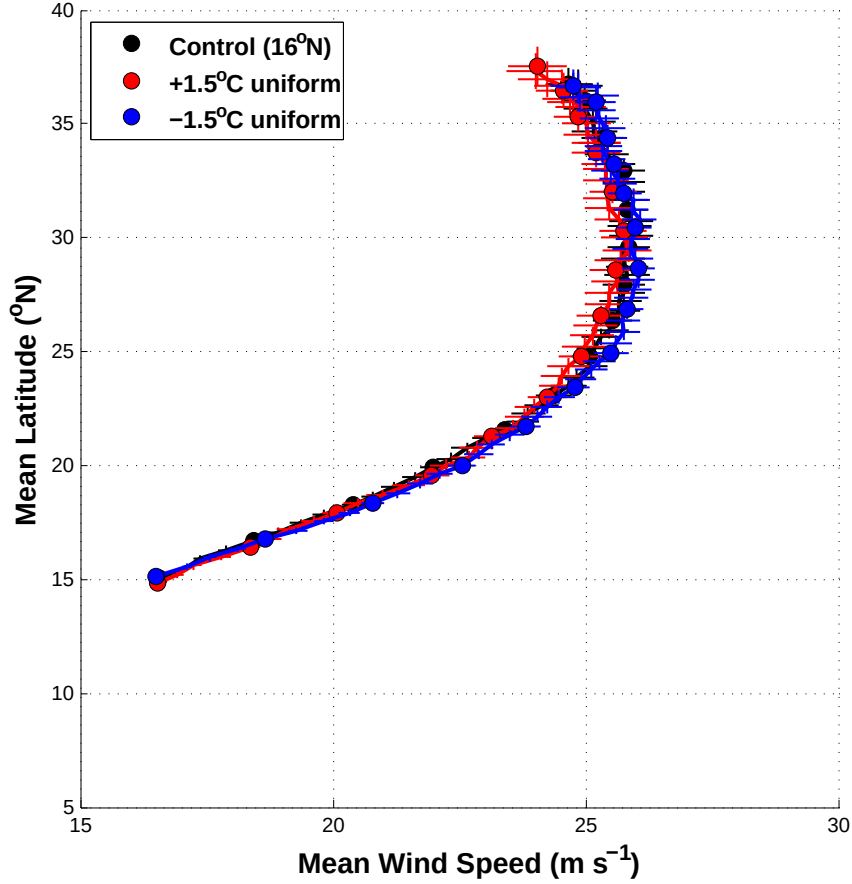


Figure 4.6: The time evolution of TCs in the HiRAM aquaplanet fixed-SST simulations with a $\pm 1.5^\circ\text{C}$ globally uniform SST perturbation ($\theta_0 = 16^\circ\text{N}$). Starting from genesis and plotting subsequent 6-hr values for a 14-day period, each cross shows the 5-95% bootstrap confidence interval for the estimates of the mean latitude ($^\circ\text{N}$) and mean intensity (surface wind speed, m s^{-1}), as a function of track duration. Markers of the mean latitude and intensity are plotted every 1 day of track duration.

Evolution of the TC environment

The globally uniform perturbation to the underlying SST is immediately apparent in the respective differences plotted in the evolution of the TC environment (Fig. 4.9). In particular, both the SSTs (Fig. 4.9a,e) and PI (Fig. 4.9b,f) are significantly greater (less) throughout the lifetime of TCs for the case of the uniform warming (cooling) simulations.

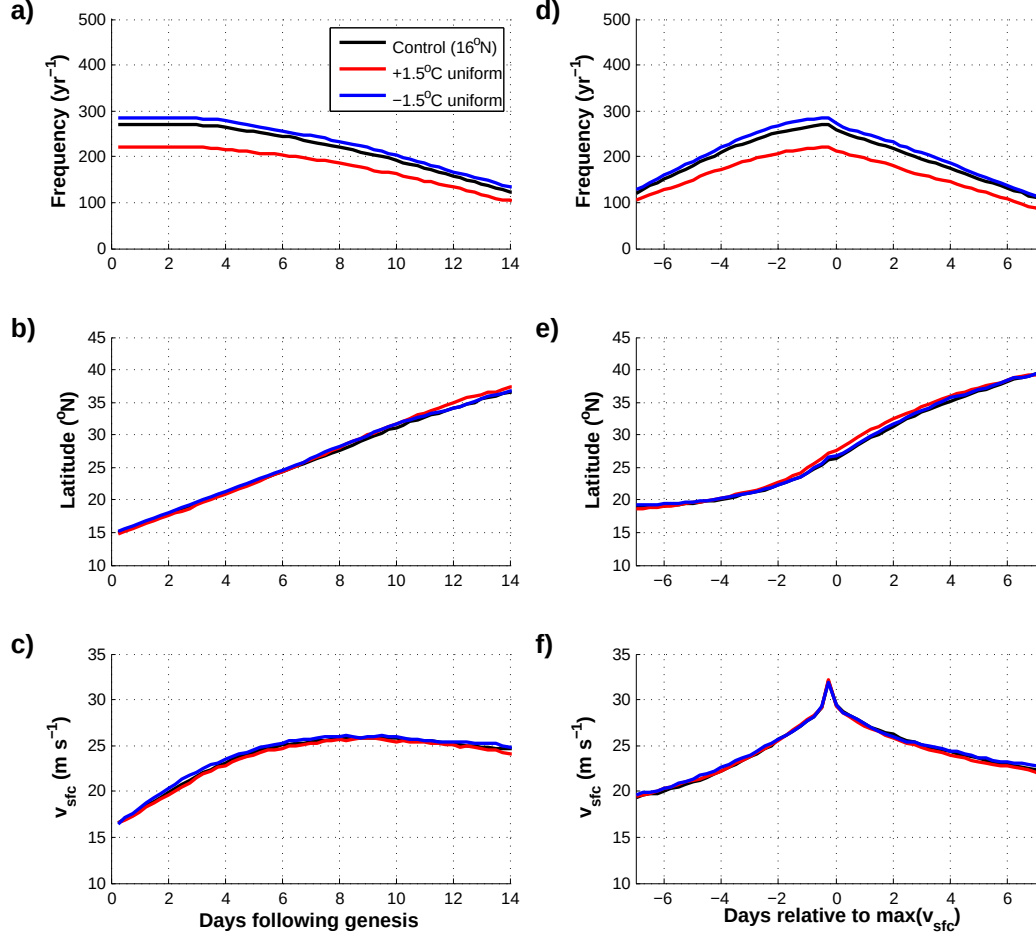


Figure 4.7: The time evolution of TCs in the HiRAM aquaplanet fixed-SST control simulations with a $\pm 1.5^\circ\text{C}$ globally uniform SST perturbation ($\theta_0 = 16^\circ\text{N}$), with reference to the time of first detection [a,b,c], and to the time of the lifetime maximum intensity [d,e,f]. Plotted is the frequency (yr^{-1}) [a,d]; the mean latitude ($^\circ\text{N}$) [b,e]; and the mean intensity (m s^{-1}) [c,f] of the tracked disturbances.

The differences in vertical motion (Fig. 4.9c,g) are slighter. While the global warming scenario yields weaker mean vertical motion during the period of TC intensification, no noticeable change (in the mean) is seen in the TC environment of the global cooling simulation. This nonlinearity (in the response to a warming versus cooling) was noted with respect to differences in the strength of the ITCZ (Fig. 3.4b) and corresponding rate of TC genesis (Tab. 3.1).

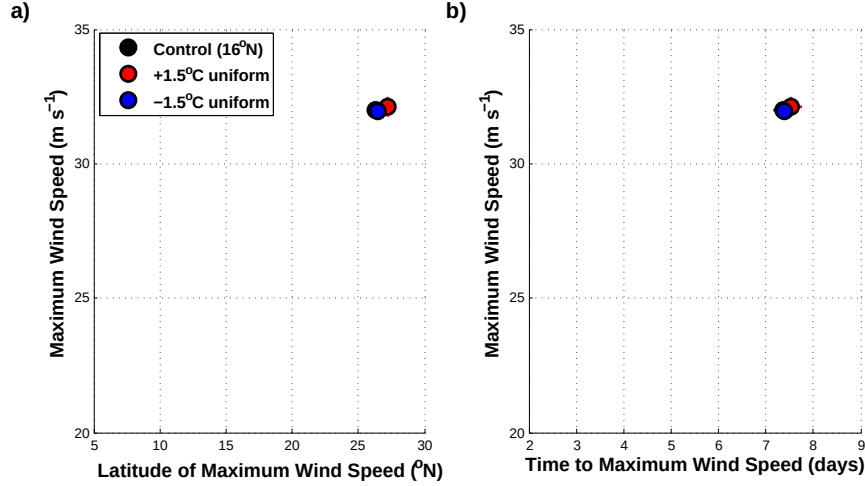


Figure 4.8: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet fixed-SST simulations with a $\pm 1.5^\circ \text{C}$ globally uniform SST perturbation ($\theta_0 = 16^\circ \text{N}$), plotted against the a) mean latitude corresponding to where the TCs achieve their maximum wind speed ($^\circ \text{N}$); and b) the time interval between genesis and when the TCs achieve their maximum wind speed (days). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

The conditions of the large-scale environment coincident with the point of TC lifetime maximum intensity are summarized in Figure 4.10. With a globally uniform warming or cooling TCs achieve a similar average lifetime maximum intensity. They do so in an environment characterized by different values of SST (Fig. 4.10a), PI (Fig. 4.10b) and vertical wind shear (Fig. 4.10d). As with the poleward shift experiments, the vertical velocity (Fig. 4.10) is close to zero (moving from mean ascent to mean descent) in all experiments around the point that the maximum intensity is acquired.

Summary

When a globally uniform 1.5°C warming (or cooling) is applied to the SST in the HiRAM aquaplanet fixed-SST control simulation:

- The mean maximum intensity of simulated TCs remains the same.

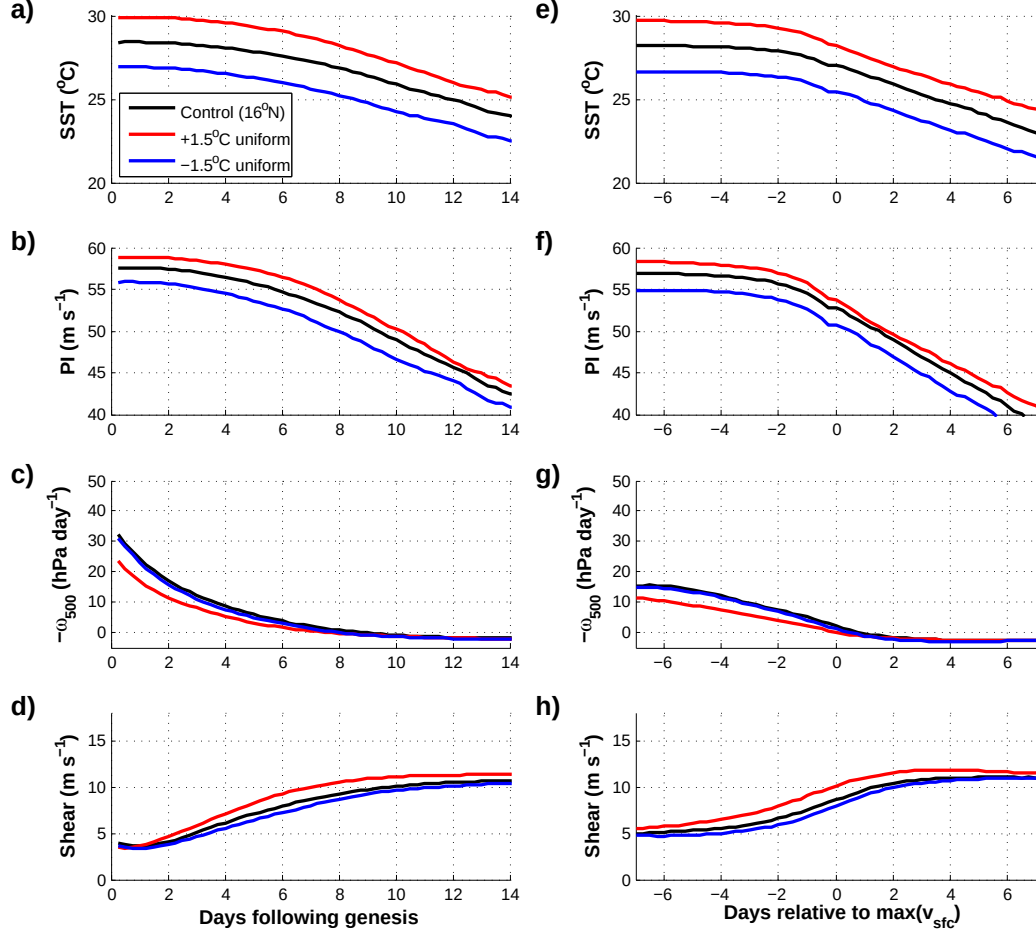


Figure 4.9: The time evolution of the the mean environmental conditions that the simulated TCs traverse throughout their lifetime, with reference to the time of first detection [a,b,c,d], and to the time of their lifetime maximum intensity [e,f,g,h]. Plotted is the SST ($^{\circ}\text{C}$) [a,e]; the Potential Intensity (m s^{-1}) [b,f]; the 500-hPa vertical pressure velocity (hPa day^{-1}) [c,g]; and the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}) [d,h].

- TCs reach their maximum intensity at a similar latitude.
- TCs have a similar-length period of intensification.
- TCs experience an environment where the underlying SSTs are significantly warmer (cooler), the PI is greater (less), the mean vertical velocity is slightly weaker, and the vertical wind shear is slightly stronger (weaker).

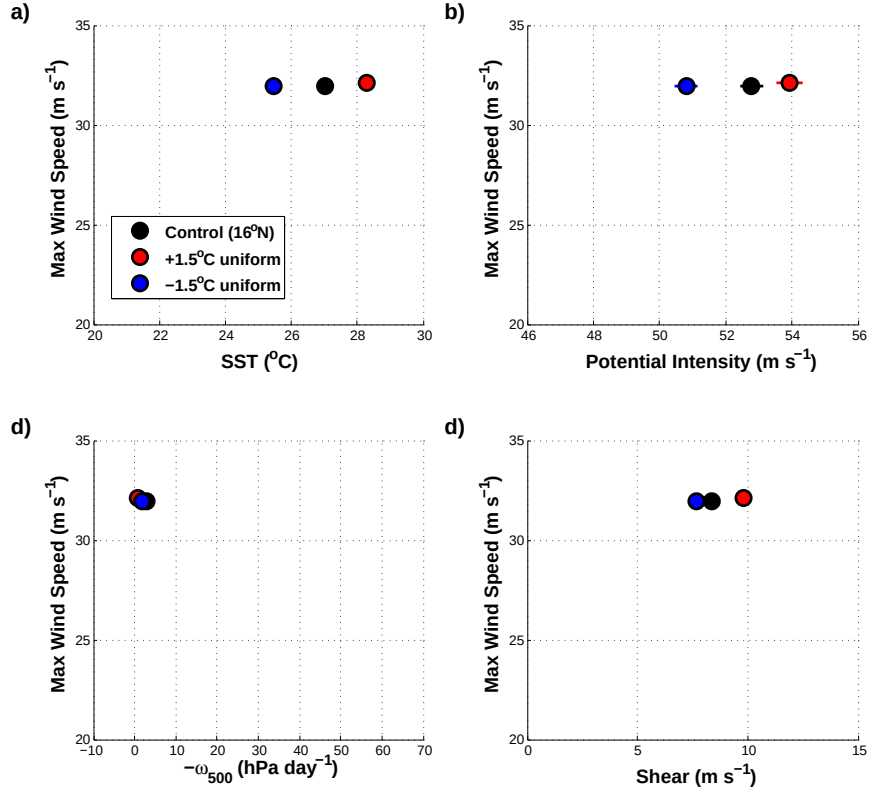


Figure 4.10: The average lifetime maximum intensity of TCs in the HiRAM aqua-planet fixed-SST simulations with a $\pm 1.5^\circ\text{C}$ globally uniform SST perturbation ($\theta_0 = 16^\circ\text{N}$), plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity. Plotted is the a) SST ($^\circ\text{C}$); b) Potential Intensity (m s^{-1}); c) 500-hPa vertical pressure velocity (hPa day^{-1}); and d) the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

4.2.3 Tropical perturbation experiments

Evolution of TC intensity

To investigate the impact of a tropical SST perturbation and associated change in meridional SST gradient, the mean time evolution of all TCs in the three control simulations ($\theta_0 = \{10^\circ\text{N}, 13^\circ\text{N}, 16^\circ\text{N}\}$), along with the six simulations corresponding to the warm and cool tropical perturbations ($\pm 1.5^\circ\text{C}$), are shown in Figure 4.11.

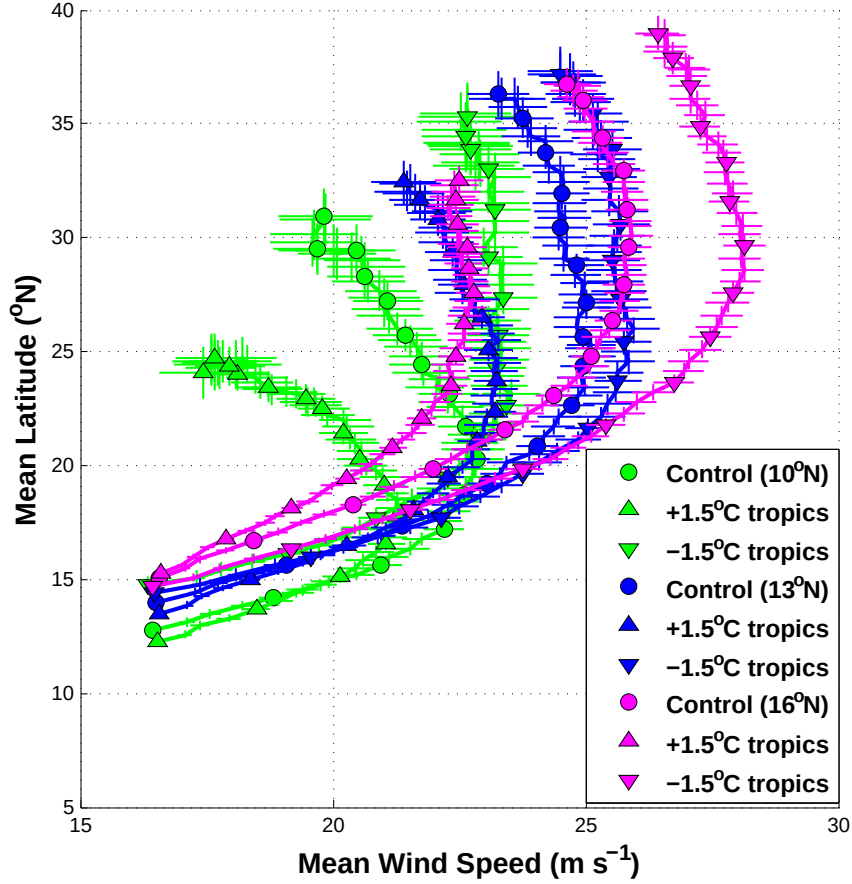


Figure 4.11: The time evolution of TCs in the HiRAM aquaplanet fixed-SST control simulations with a $\pm 1.5^\circ\text{C}$ tropical SST perturbation. Starting from genesis and plotting subsequent 6-hr values for a 14-day period, each cross shows the 5-95% bootstrap confidence interval for the estimates of the mean latitude ($^\circ\text{N}$) and mean intensity (surface wind speed, m s^{-1}), as a function of track duration. Markers of the mean latitude and intensity are plotted every 1 day of track duration.

The mean genesis latitude shifts poleward as the maximum in SST shifts poleward in these simulations, although the strength of this relationship depends on the sharpness of the meridional SST profile. When considering the entire suite of HiRAM aquaplanet experiments it was previously noted (Sec. 3.5) that the *preferred* (or *default*) genesis latitude in the limit of vanishing meridional SST gradient is $\sim 14.5^\circ\text{N}$. One can consider the actual mean genesis latitude realized in the individual simulations as being shifted (from 14.5°N) towards the latitude of the maximum SST by an amount that is related to the strength of the prescribed meridional SST gradient.

This is evident when looking at the beginning of the mean TC traces in Figure 4.11. The mean genesis latitudes in the simulations with a $+1.5^\circ\text{C}$ tropical perturbation are closer to the latitudes corresponding to the maximum in SST (10°N , 13°N , 16°N), when compared with the genesis latitudes of the simulations with a -1.5°C tropical perturbation, which remain at $\sim 14.5^\circ\text{N}$ as the maximum in SST moves poleward.

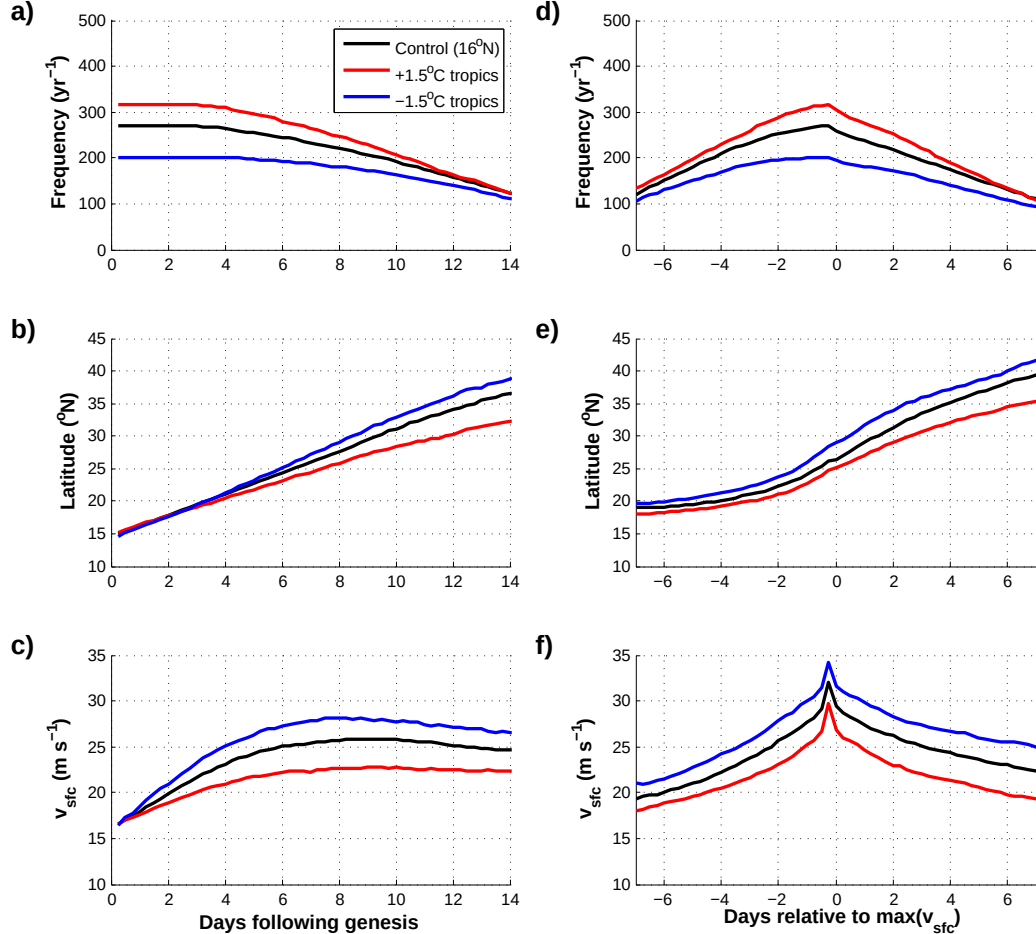


Figure 4.12: The time evolution of TCs in the HiRAM aquaplanet fixed-SST control simulations with a $\pm 1.5^\circ\text{C}$ tropical SST perturbation ($\theta_0 = 16^\circ\text{N}$), with reference to the time of first detection [a,b,c], and to the time of the lifetime maximum intensity [d,e,f]. Plotted is the frequency (yr^{-1}) [a,d]; the mean latitude ($^\circ\text{N}$) [b,e]; and the mean intensity (m s^{-1}) [c,f] of the tracked disturbances.

Figure 4.12 shows the evolution of frequency (a), latitude (b), and intensity (c) of the TCs with reference to the genesis time (left panels) for the $\theta_0 = 16^\circ\text{N}$ tropical

perturbation experiments. The increase (decrease) in the simulated genesis frequency (Fig. 4.12a) associated with a tropical warming (cooling) was discussed in Section 3.4. Note also here from the shape of these frequency (of TC duration) profiles that the TCs in the warmer (cooler) simulations have a relatively shorter (longer) average lifetime duration than those in the control simulations. The initial marked difference in genesis frequency reduces substantially (to almost zero) after fourteen days.

Although the TCs in the $\theta_0 = 16^\circ \text{N}$ tropical perturbation experiments have the same mean genesis latitude ($\sim 14.5^\circ \text{N}$), the lines plotting the subsequent evolution of TC latitude (Fig. 4.12b) show that the TCs in the different simulations move polewards at different rates. TCs in the tropical warming climate (with a sharper meridional SST gradient) move polewards more slowly, remaining at lower latitudes, while those TCs simulated in a flatter-SST tropical cooling climate move polewards more rapidly (on average). Furthermore, the comparison of TC intensity evolution (Fig. 4.12c) shows that the TCs which form in the cooler tropical climate develop substantially higher wind speeds than those in the tropical warming simulations.

The right panels of Figure 4.12 show the TC evolution with respect to the time of maximum intensity. The mean latitude of lifetime maximum intensity and the mean period of intensification are also plotted in Figure 4.13. TCs in the simulations having a cool (warm) tropical perturbation and thus a flatter (sharper) meridional gradient of tropical SST, on average achieve a higher (lower) lifetime maximum intensity at a mean latitude that is further poleward (equatorward) than the TCs in the control experiments (Fig. 4.12e, 4.13a), having a longer (shorter) period of time between genesis and their lifetime maximum intensity (Fig. 4.13b).

Evolution of the TC environment

The evolution of the TC environment (SST, PI, vertical velocity and vertical wind shear) in the $\theta_0 = 16^\circ \text{N}$ tropical perturbation experiments is shown in Figure 4.14,

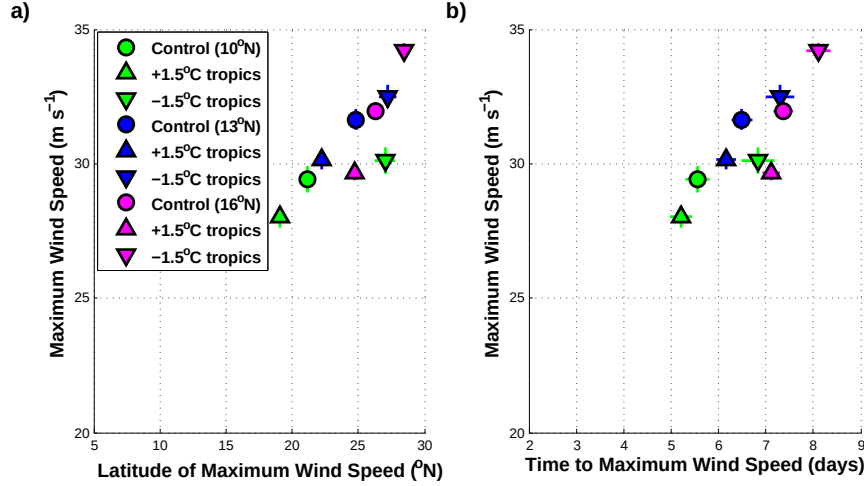


Figure 4.13: The average lifetime maximum intensity of TCs in the HiRAM aqua-planet fixed-SST simulations with a $\pm 1.5^{\circ}\text{C}$ tropical SST perturbation, plotted against the a) mean latitude corresponding to where the TCs achieve their maximum wind speed ($^{\circ}\text{N}$); and b) the time interval between genesis and when the TCs achieve their maximum wind speed (days). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

with the associated mean values (for all tropical perturbation experiments) at the time and latitude of maximum intensity shown in Figure 4.15.

The TCs in the tropical warming (cooling) simulations experience underlying SSTs (Fig. 4.14a,e) that are markedly warmer (cooler) throughout their lifetime, along with correspondingly higher (lower) values of mean PI (Fig. 4.14a,e). Hence the greater (weaker) lifetime maximum intensities realized in the cooler (warmer) perturbation simulations are associated with lower (higher) absolute values of SST (Fig. 4.15a) and PI (Fig. 4.15b).

Following genesis and for the duration of the intensification phase, TCs in the tropical perturbation simulations experience a significant difference in mean vertical velocity than in the control simulation (Fig. 4.14c,g). This is consistent with the marked difference observed in the mean ITCZ (Fig. 3.5b), where the tropical warming (cooling) is associated with substantially stronger (weaker) ascent in the ascending

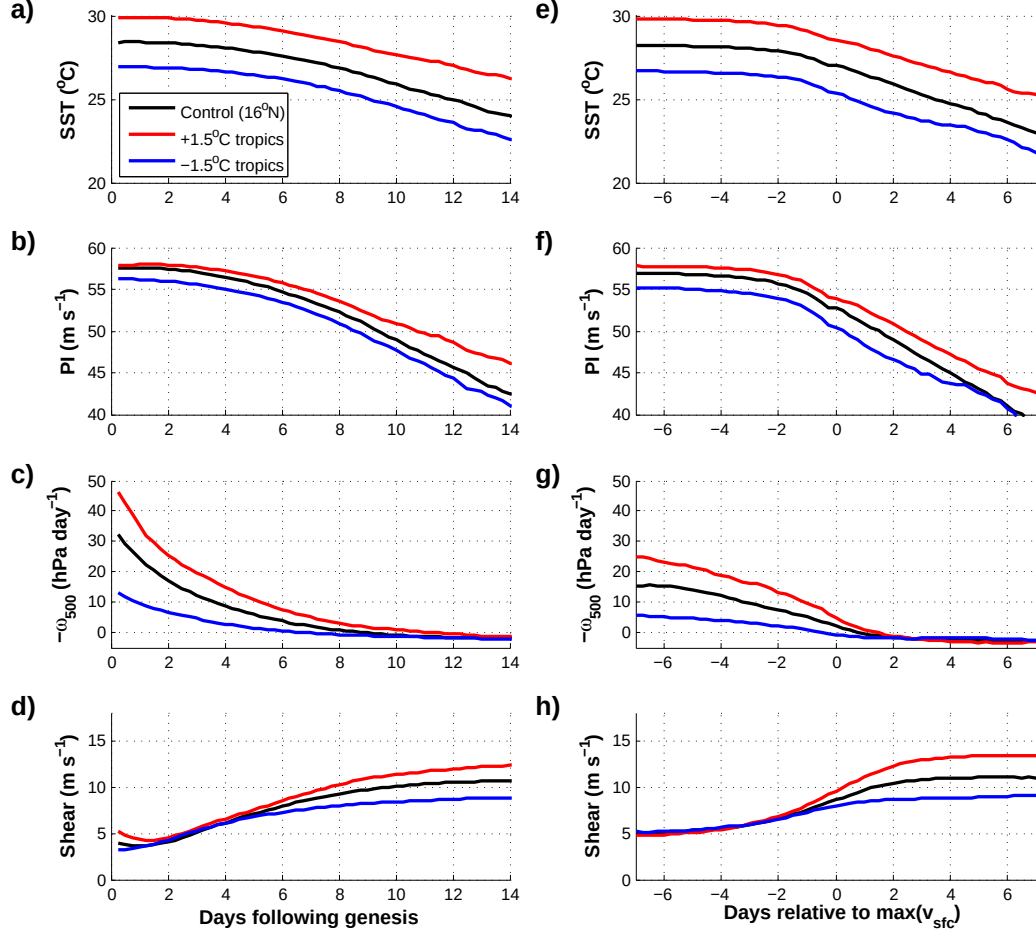


Figure 4.14: The time evolution of the the mean environmental conditions that the simulated TCs traverse throughout their lifetime, with reference to the time of first detection [a,b,c,d], and to the time of their lifetime maximum intensity [e,f,g,h]. Plotted is the SST ($^{\circ}\text{C}$) [a,e]; the Potential Intensity (m s^{-1}) [b,f]; the 500-hPa vertical pressure velocity (hPa day^{-1}) [c,g]; and the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}) [d,h].

branch (equatorward of $\sim 28^{\circ}\text{N}$). As the TCs move poleward they reach a maximum intensity as the vertical ascent reduces to zero (Fig. 4.15c),

While prior to the TCs reaching their maximum intensity the magnitude of vertical wind shear is very similar across experiments, the time evolution curves begin to diverge at about day = -1 (Fig. 4.14h). TCs in the tropical warming scenario subsequently encounter stronger vertical wind shear than in the cool perturbation

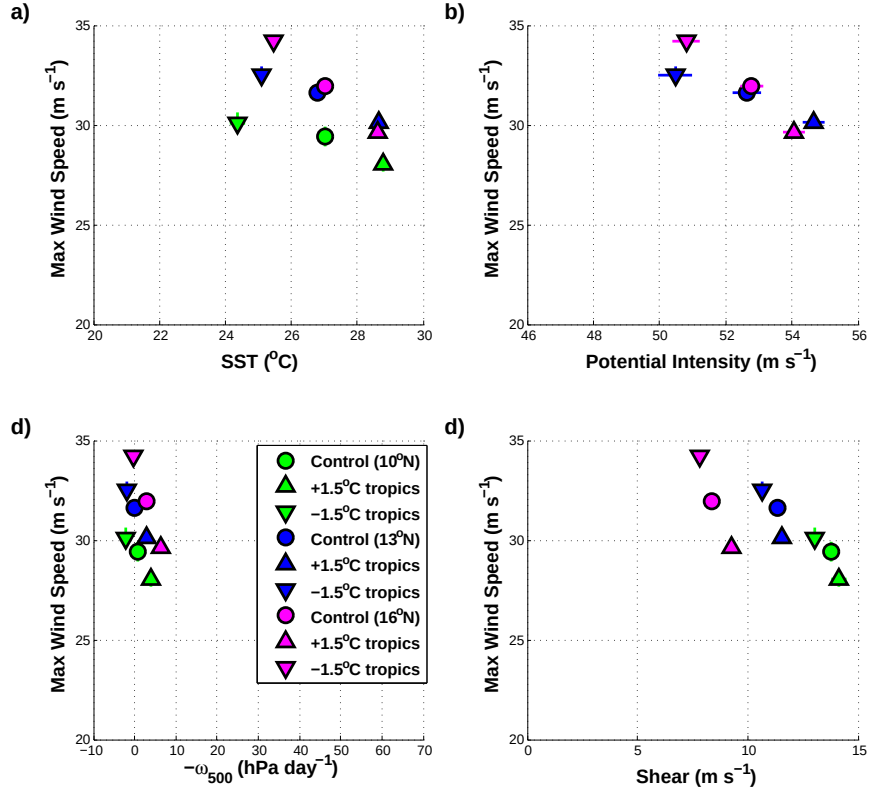


Figure 4.15: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet fixed-SST simulations with a $\pm 1.5^\circ\text{C}$ tropical SST perturbation, plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity. Plotted is the a) SST ($^\circ\text{C}$); b) Potential Intensity (m s^{-1}); c) 500-hPa vertical pressure velocity (hPa day^{-1}); and d) the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

experiments. The relationship shown in Figure 4.15d suggests one of decreasing maximum lifetime intensity at higher values of vertical wind shear.

Summary

When a tropically localized 1.5°C warming (or cooling) is applied to the SST in the HiRAM aquaplanet fixed-SST control simulation:

- The mean maximum intensity of simulated TCs increases (decreases).

- TCs reach their maximum intensity at a higher (lower) latitude and have a longer (shorter) period of intensification.
- TCs experience an environment where the underlying SSTs are significantly warmer (cooler) with an associated increase (decrease) in PI.
- During the intensification phase, TCs experience a large-scale environment characterized by stronger (weaker) mean vertical velocity, but a very similar magnitude of vertical wind shear.
- As TCs reach their maximum intensity and enter the weakening phase, the vertical wind shear becomes comparatively stronger (weaker), and the differences in vertical velocity are no longer evident.

4.2.4 Slab ocean experiments

Evolution of TC intensity

The mean time evolution of all TCs simulated in the HiRAM aquaplanet slab ocean configuration are shown in Figure 4.16. Results are shown for simulations with increasing inter-hemispheric asymmetry, $Q_0 = \{20, 40, 50, 60\} \text{ Wm}^{-2}$ (square markers).

The evolution of the TC statistics (duration frequency, latitude, and intensity) following genesis are shown in Figure 4.17 (left panels). The mean latitude of TC genesis shifts poleward as the asymmetric forcing increases (Fig. 4.16, and recall Sec. 3.5). This difference in mean latitude (between the slab ocean simulations) increases with time following genesis (Fig. 4.17b); in two weeks the TCs in the $Q_0 = 40 \text{ Wm}^{-2}$ simulation move from $\sim 10\text{-}15^\circ \text{ N}$, whereas in the $Q_0 = 60 \text{ Wm}^{-2}$ simulation they move from $\sim 16\text{-}28^\circ \text{ N}$.

Along with the difference in poleward movement, the TCs in the slab ocean simulations acquire (on average) a different mean lifetime intensity (Fig. 4.17c,f). Figure

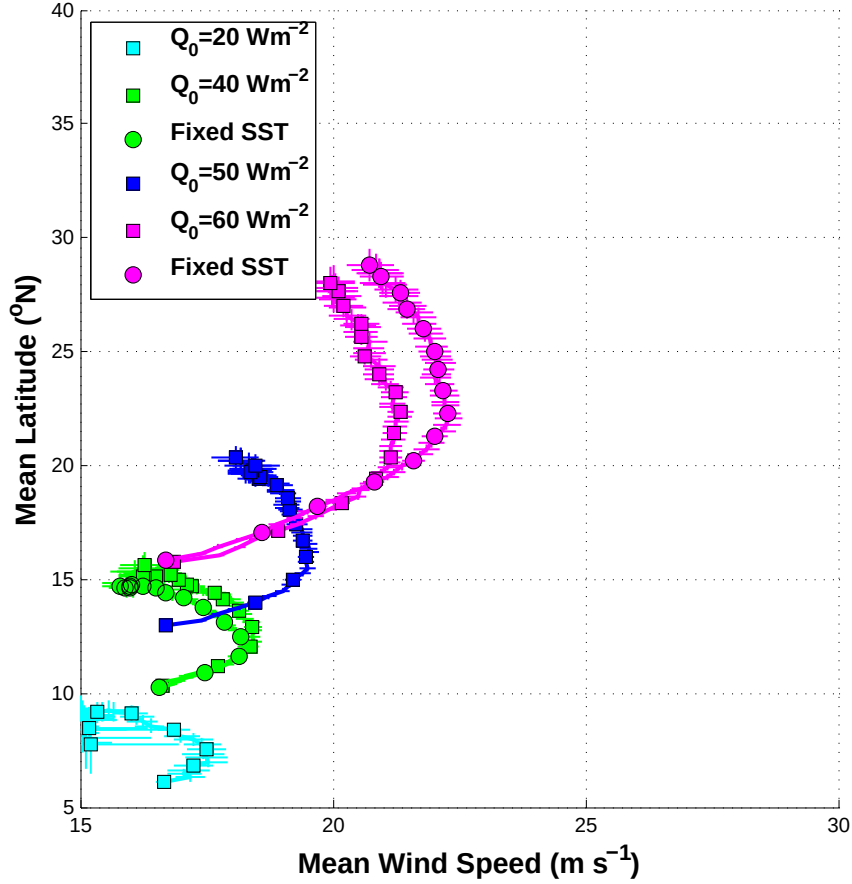


Figure 4.16: The time evolution of TCs in the HiRAM aquaplanet simulations in the slab ocean configuration ($Q_0 = \{20, 40, 50, 60\} \text{ Wm}^{-2}$). Two fixed SST simulations ($Q_0 = \{40, 60\} \text{ Wm}^{-2}$, circle markers) are also shown for comparison, where the SST climatology previously diagnosed from the slab ocean has been prescribed as the lower boundary condition. Starting from genesis and plotting subsequent 6-hr values for a 14-day period, each cross shows the 5-95% bootstrap confidence interval for the estimates of the mean latitude ($^{\circ} \text{N}$) and mean intensity (surface wind speed, m s^{-1}), as a function of track duration. Markers of the mean latitude and intensity are plotted every 1 day of track duration.

4.18 shows a strongly correlated relationship between the latitude and magnitude of lifetime maximum intensity, along with the period of intensification. As the inter-hemispheric asymmetry increases the TCs experience an increasingly longer period of intensification (Fig. 4.18b), achieving (on average) a greater lifetime maximum intensity at a higher mean latitude (Fig. 4.18a).

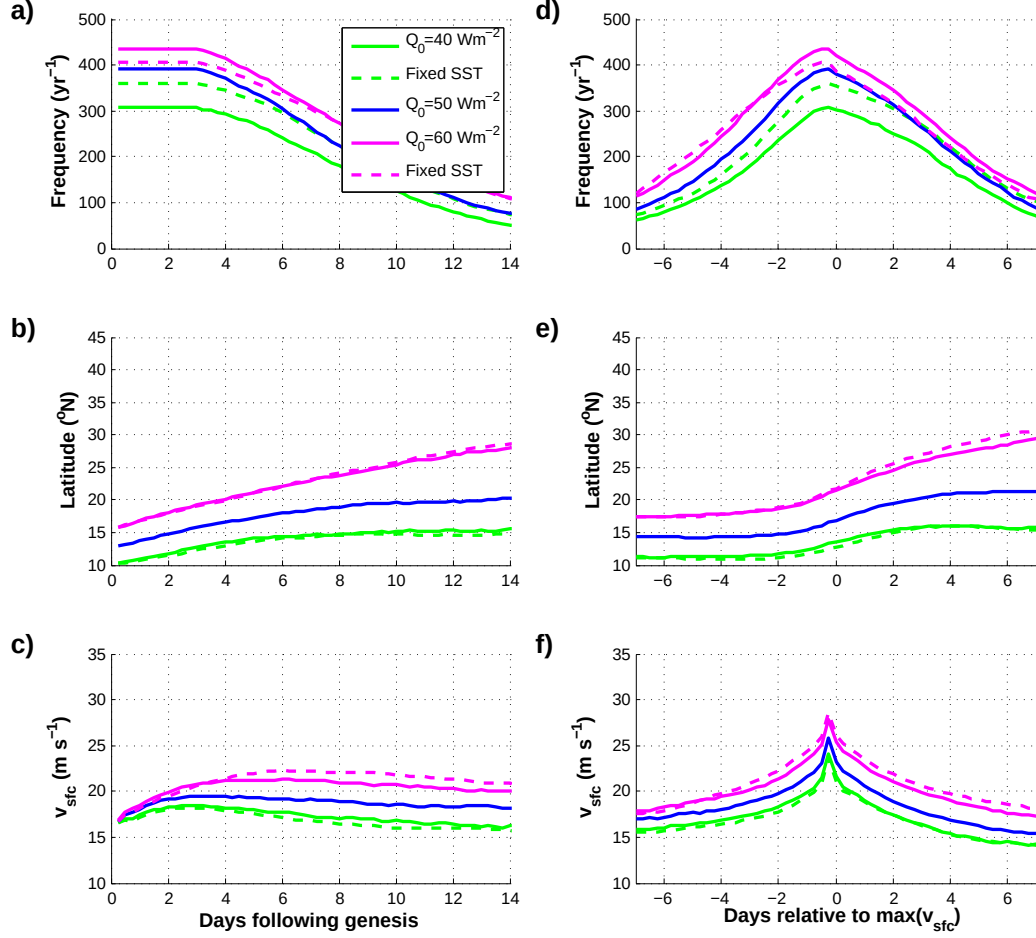


Figure 4.17: The time evolution of TCs in the HiRAM aquaplanet simulations in the slab ocean configuration ($Q_0 = \{40, 50, 60\} \text{ Wm}^{-2}$), with reference to the time of first detection [a,b,c], and to the time of the lifetime maximum intensity [d,e,f]. Plotted is the frequency (yr^{-1}) [a,d]; the mean latitude ($^{\circ} \text{N}$) [b,e]; and the mean intensity (m s^{-1}) [c,f] of the tracked disturbances.

Two fixed-SST simulations ($Q_0 = \{40, 60\} \text{ Wm}^{-2}$, circle markers) have also been plotted for comparison. For these simulations the SST climatology previously diagnosed from the slab ocean has been prescribed as the lower boundary condition. Figures 4.16 and 4.17 show that the impact of fixing the SSTs on the TC statistics differs between the $Q_0 = 40 \text{ Wm}^{-2}$ and $Q_0 = 60 \text{ Wm}^{-2}$ scenarios. For the $Q_0 = 40 \text{ Wm}^{-2}$ scenario comparison, fixing the SSTs yields a greater number of TCs that have a slightly lower lifetime maximum intensity and migrate slightly less poleward than in

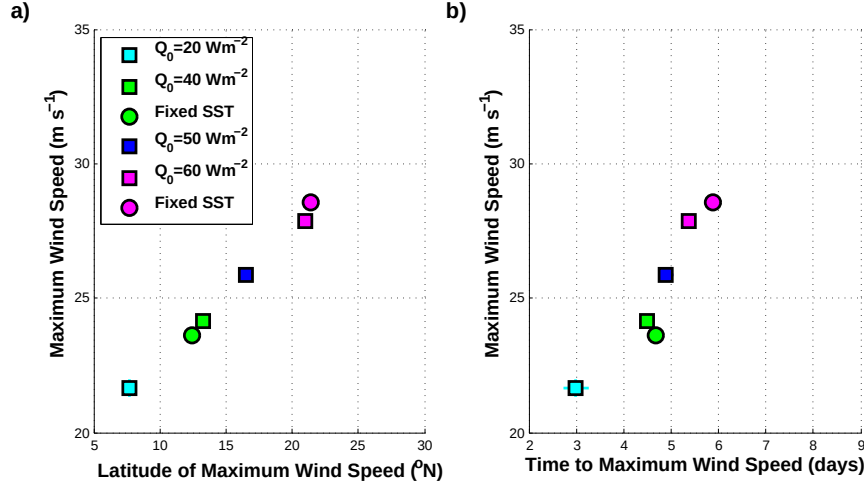


Figure 4.18: The average lifetime maximum intensity of TCs in the HiRAM aqua-planet simulations in the slab ocean configuration ($Q_0 = \{20, 40, 50, 60\} \text{ Wm}^{-2}$), plotted against the a) mean latitude corresponding to where the TCs achieve their maximum wind speed ($^{\circ}\text{N}$); and b) the time interval between genesis and when the TCs achieve their maximum wind speed (days). Two fixed-SST simulations ($Q_0 = \{40, 60\} \text{ Wm}^{-2}$, circle markers) are also shown for comparison, where the SST climatology previously diagnosed from the slab ocean has been prescribed as the lower boundary condition. Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

the slab ocean simulations. In the case of the $Q_0 = 60 \text{ Wm}^{-2}$ scenario comparison, fixing the SSTs yields fewer TCs, but they have a slightly higher lifetime maximum intensity, and migrate slightly further poleward.

Evolution of the TC environment

The evolution of the TC environment post-genesis is characterized by decreasing SSTs, Potential Intensity, and vertical motion, along with increasing vertical wind shear (Fig. 4.19). As the inter-hemispheric asymmetrical heating increases (between the different simulations) the TCs experience relatively warmer SSTs (Fig. 4.19a), higher PI (Fig. 4.19b), and weaker vertical ascent (Fig. 4.19c) as they intensify following genesis. The maximum in lifetime intensity coincides with a sharp decrease

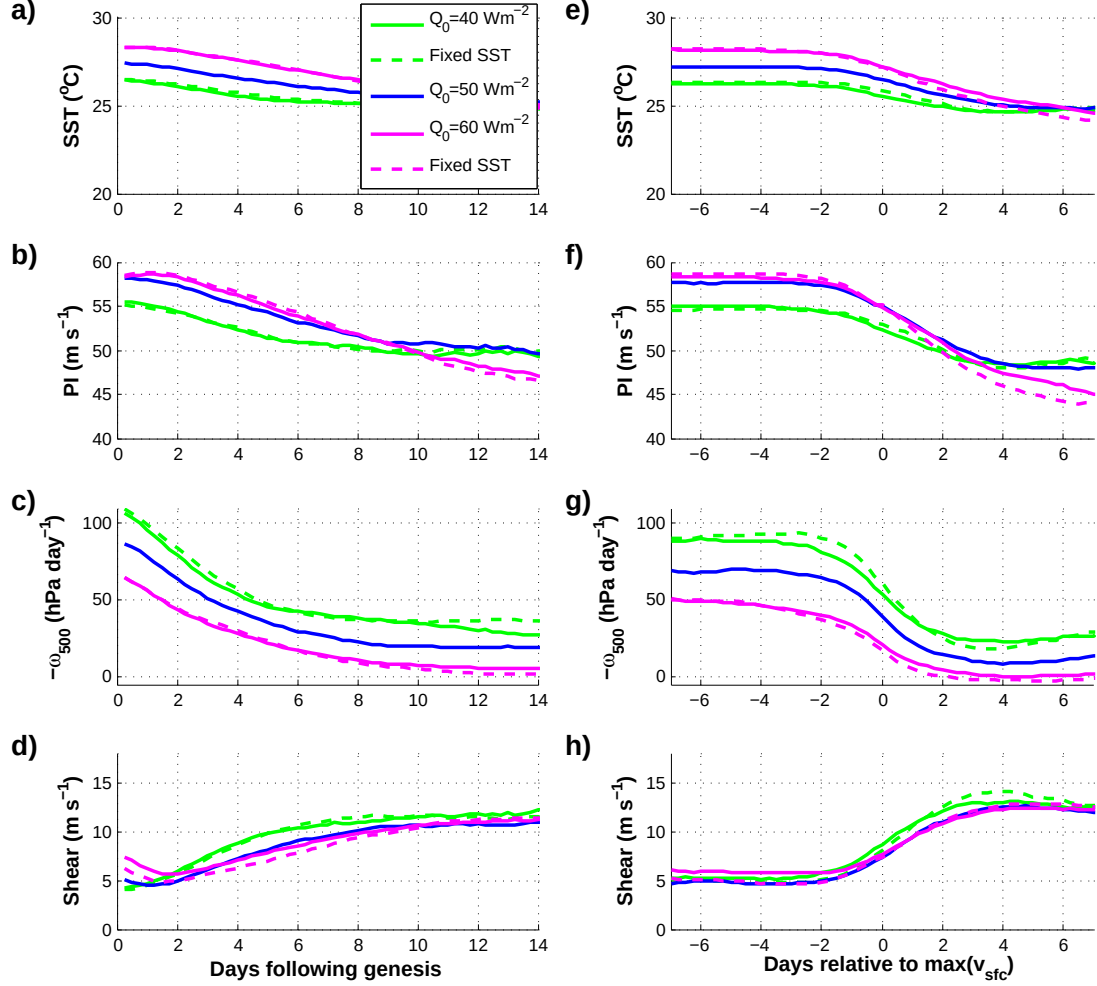


Figure 4.19: The time evolution of the the mean environmental conditions that the simulated TCs traverse throughout their lifetime, with reference to the time of first detection [a,b,c,d], and to the time of their lifetime maximum intensity [e,f,g,h]. Plotted is the SST ($^{\circ}\text{C}$) [a,e]; the Potential Intensity (m s^{-1}) [b,f]; the 500-hPa vertical pressure velocity (hPa day^{-1}) [c,g]; and the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}) [d,h].

in PI (Fig. 4.19f) and vertical ascent (Fig. 4.19g), and a marked increase in the magnitude of vertical wind shear (Fig. 4.19h).

The mean values of the large-scale environment coincident with the time when the simulated TCs reach their lifetime maximum intensity are plotted in Figure 4.20. Within this subset of HiRAM simulations (in the slab ocean configuration) there

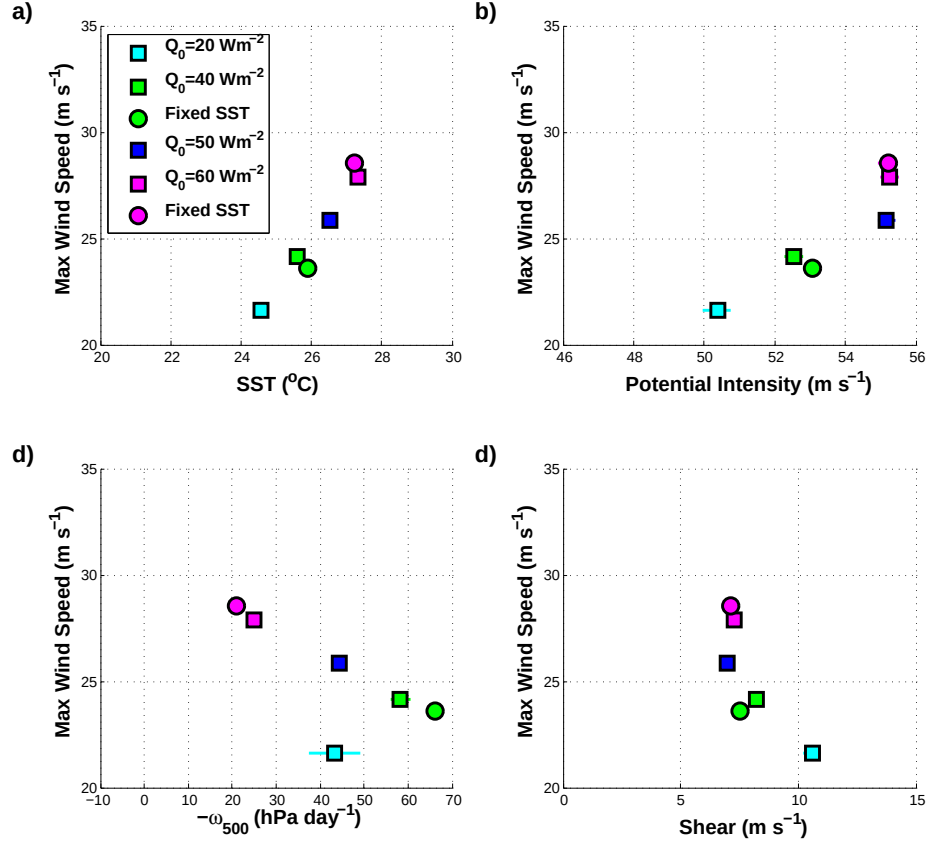


Figure 4.20: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet simulations in the slab ocean configuration ($Q_0 = \{20, 40, 50, 60\} \text{ Wm}^{-2}$), plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity. Plotted is the a) SST ($^{\circ}\text{C}$); b) Potential Intensity (m s^{-1}); c) 500-hPa vertical pressure velocity (hPa day^{-1}); and d) the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

appear to emerge some robust relationships between the large-scale environment at the time of maximum intensity and the maximum intensity itself.

Firstly, the maximum intensity is positively correlated with the underlying SST (Fig. 4.20a), and likewise with the Potential Intensity (Fig. 4.20b). Somewhat surprisingly, the lifetime maximum intensity appears to be anti-correlated with the strength of the coincident vertical motion (Fig. 4.20c). There is no clear relationship in the different values of vertical wind shear (Fig. 4.20d).

Summary

As the inter-hemispheric asymmetry increases:

- The mean maximum intensity of simulated TCs increases.
- TCs reach their maximum intensity at higher latitudes.
- TCs have a longer period of intensification, reaching a maximum intensity later in their lifetime.
- TCs experience an environment where the underlying SSTs, the Potential Intensity, and the vertical velocity are all higher. The magnitude of the vertical wind shear remains similar.
- Despite the differing behavior of the $Q_0 = 40 \text{ Wm}^{-2}$ and $Q_0 = 60 \text{ Wm}^{-2}$ fixed-SST simulations, the general result holds that greater mean maximum intensities occur at higher latitudes across all of the simulations.

4.3 Summary

TCs form on the poleward flank of the ITCZ and migrate to higher latitudes, strengthening for a while before reaching their maximum intensity and decaying. The evolution of the TC environment during this post-genesis poleward migration is characterized by decreasing SSTs (as the lower boundary condition dictates), decreasing Potential Intensity (closely tied to the decreasing SSTs), decreasing vertical motion (moving away from the ITCZ region into the subsidence of the extra-tropics), and increasing vertical wind shear. The average lifetime maximum intensity of TCs in the HiRAM aquaplanet simulations have been plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity in Figure 4.21.

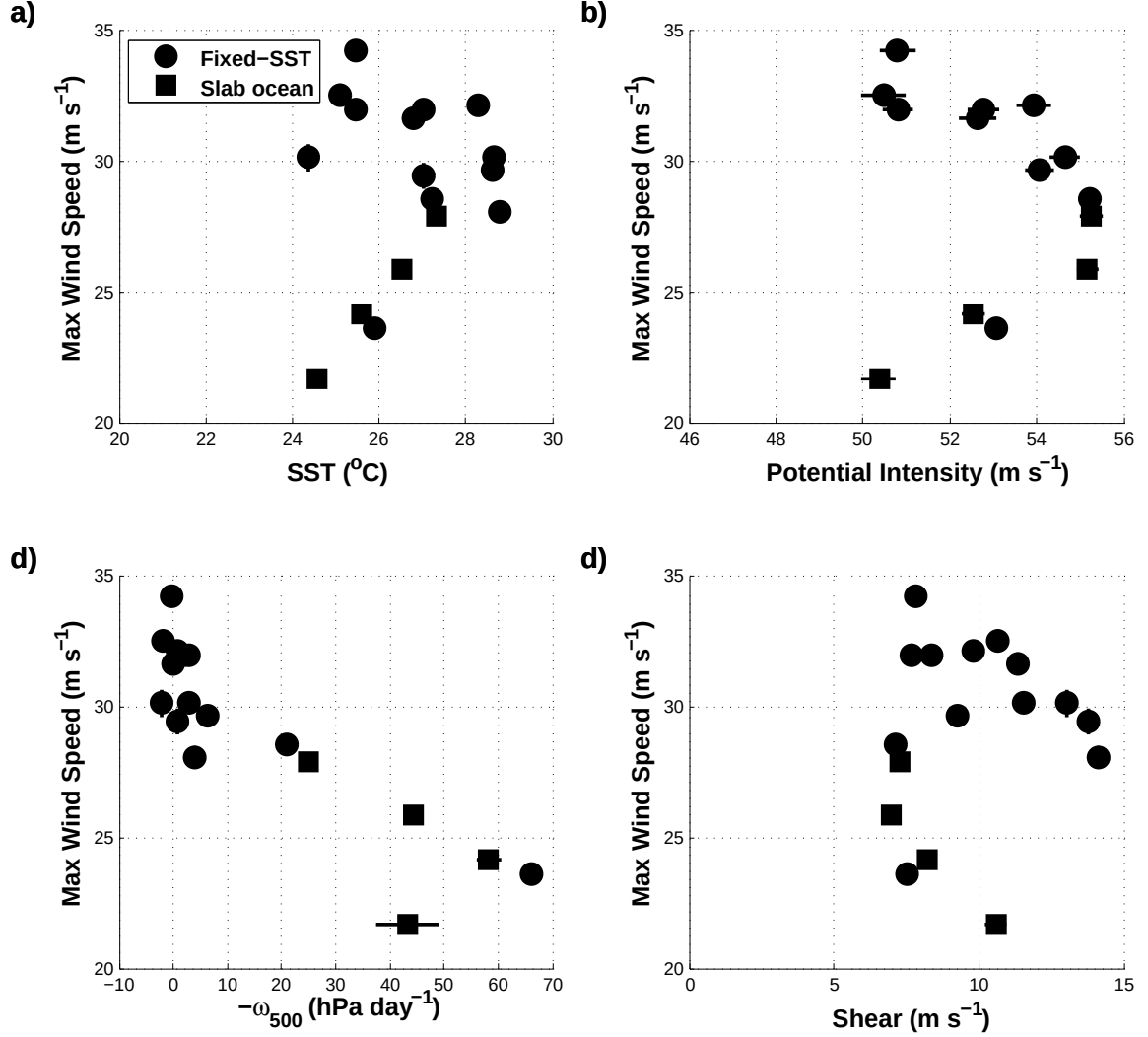


Figure 4.21: The average lifetime maximum intensity of TCs in the HiRAM aquaplanet simulations, plotted against the mean environmental conditions experienced by the TCs at the time and latitude of their maximum intensity. Plotted is the a) SST ($^{\circ}\text{C}$); b) Potential Intensity (m s^{-1}); c) 500-hPa vertical pressure velocity (hPa day^{-1}); and d) the magnitude of the vertical shear (250-850 hPa) of the horizontal wind (m s^{-1}). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

Mean latitude of TC maximum intensity

Summarizing the results from this section, Figure 4.22 shows the robust result relating lifetime maximum wind speed to the latitude at which the maximum is reached.

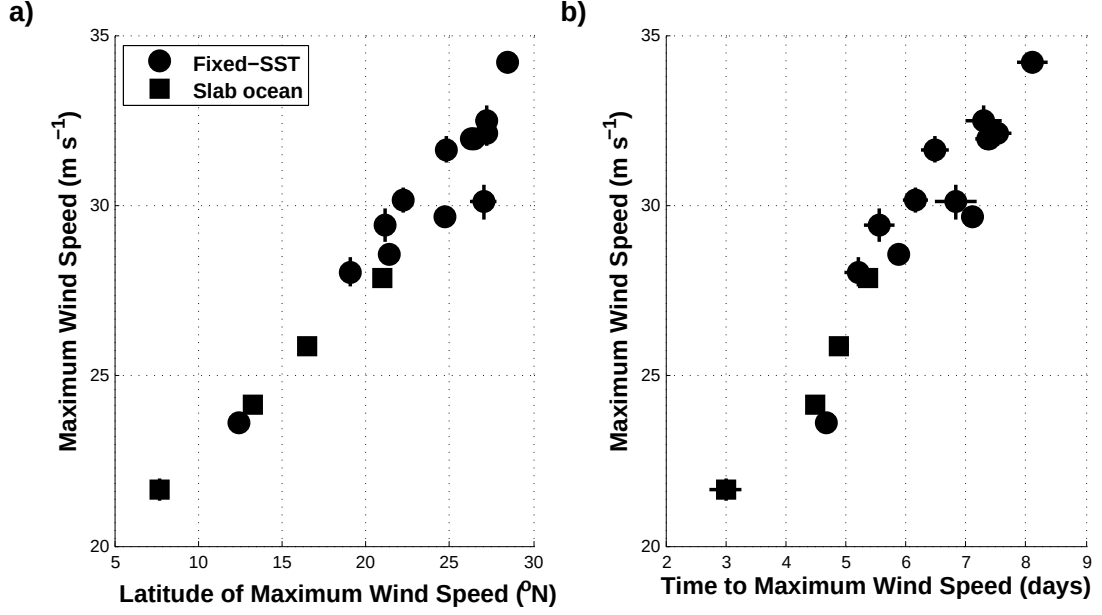


Figure 4.22: The average lifetime maximum intensity of TCs across all of the HiRAM aquaplanet simulations, plotted against the a) mean latitude corresponding to where the TCs achieve their maximum wind speed ($^{\circ}\text{N}$); and b) the time interval between genesis and when the TCs achieve their maximum wind speed (days). Crosses indicate the 5-95% bootstrap confidence interval for the estimates of the mean.

Across a diverse range of aquaplanet experiments these plots show a positive relationship between the mean lifetime maximum intensity of TCs and the corresponding mean latitude at which the maximum occurs (Fig. 4.22a). There is also a strong correlation between the mean lifetime maximum intensity of TCs and the period of time that the TCs spend intensifying (Fig. 4.22b). These two aspects of the TC tracks (the latitude of maximum intensity, and the period until maximum intensity) are closely related, and it would be interesting to investigate this inter-dependence further.

Chapter 5

Zonally asymmetric forcing

5.1 Introducing the perturbation experiments

Up until this point we have only considered zonally-symmetric climates of the HiRAM aquaplanet and the differences in the simulated tropical cyclone activity associated with relatively small changes in the zonally-symmetric SST boundary condition. We have explored the poleward shift of the underlying SST profile, along with globally uniform SST perturbations or perturbations to the tropical meridional SST gradient, all whilst maintaining zonal symmetry through fixing the prescribed SST. In the case of the simulations in the slab ocean configuration the SST was not held fixed, but the time-mean SST (later diagnosed) was nonetheless zonally-symmetric due to the nature of the forcing.

We now expand the set of fixed-SST aquaplanet experiments to include prescribed zonally-asymmetric perturbations to the zonal SST gradient. Exploring idealized warm and cool asymmetries is a natural step towards understanding the impact of large-scale basin-to-basin SST variability on global TC activity. It is motivated in part by the apparent utility of relative SST (the local SST relative to the zonal mean) in moderating tropical cyclone activity (e.g. Vecchi et al., 2008).

A series of fixed-SST experiments are conducted where a zonal asymmetry is introduced to the tropical perturbation experiments described in Section 2.1. The asymmetry is a simple cosine function of longitude λ (degrees),

$$SST'(\lambda, \theta) = SST'(\theta)|_{\text{sym}} \cos \left[\frac{\pi k}{180} (\lambda - 180) \right] \quad (5.1)$$

where $SST'(\theta)|_{\text{sym}}$ is defined by Eq. 2.2 and the wavenumber k sets the number of zonal SST maxima (and minima). Since the wave perturbation is applied as a further modulation to the symmetric tropical perturbation, the meridional extent of the asymmetric perturbation in SST is similarly confined to tropical latitudes. The perturbation is greatest at θ_0 , reducing to zero at $\theta_0 \pm 30^\circ$ (and beyond). At the longitudes corresponding to SST maxima the meridional SST profile will be identical to the warm tropical perturbation experiments explored in the previous chapters. Similarly, at the longitudes of SST minima the meridional SST profile is equivalent to the cool tropical perturbation experiments. These zonally asymmetric perturbation experiments can thus be considered a forcing that smoothly alternates (in longitude) between the warm and cold tropical perturbations considered in the previous chapters.

Along with the alternating perturbations to the meridional SST gradient, the prescribed zonal asymmetries allow an exploration of the impact of zonal SST gradients. The wavenumber sets the number of zonal SST maxima and thus also varies the mean magnitude of the zonal SST gradient. The average (over all longitudes) magnitude of the zonal gradient of SST scales linearly with wavenumber, $|\overline{\nabla_\lambda(SST)}| \propto k$.

Nine asymmetric perturbation experiments were conducted, varying $k = 1, 2, 3$ at each of $\theta_0 = \{10^\circ \text{ N}, 13^\circ \text{ N}, 16^\circ \text{ N}\}$, all with $SST_0 = 28.5^\circ \text{ C}$ and $SST'_0 = 1.5^\circ \text{ C}$ as in the symmetric experiments. The simulations were each run for a period of 5 yr, and tropical cyclone activity was subsequently analyzed using the vortex tracker previously described in Section 2.3.

5.2 Results

5.2.1 Frequency of TC genesis

A summary of the genesis frequency results for the asymmetric SST perturbation experiments is provided in Table 5.1. Results are shown for all TCs and for the subset of TCs that acquire hurricane-strength surface wind speeds during their lifetime ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$). In order to reduce the number of TCs of non-tropical origin influencing the statistics, only those which form equatorward of 30° N are included in this analysis. The main results discussed throughout the remainder of this section are relatively insensitive to the application of this genesis latitude criterion.

Configuration	Experiment	Length of simulation	Frequency (yr^{-1})	
			<i>All TCs</i>	<i>Strong TCs</i>
Fixed-SST	$\theta_0 = 10^\circ \text{ N}$ (control)	10 yr	134.8 ± 2.5	56.7 ± 1.5
	$\theta_0 = 13^\circ \text{ N}$ (control)	10 yr	203.5 ± 3.5	118.3 ± 1.5
	$\theta_0 = 16^\circ \text{ N}$ (control)	10 yr	286.1 ± 3.9	179.0 ± 2.3
	$\theta_0 = 10^\circ \text{ N}$ ($k = 1$)	5 yr	227.6 ± 2.6	98.9 ± 1.9
	$\theta_0 = 13^\circ \text{ N}$ ($k = 1$)	5 yr	259.4 ± 6.1	116.8 ± 4.9
	$\theta_0 = 16^\circ \text{ N}$ ($k = 1$)	5 yr	277.0 ± 3.7	138.6 ± 1.4
	$\theta_0 = 10^\circ \text{ N}$ ($k = 2$)	5 yr	194.7 ± 7.5	96.3 ± 3.2
	$\theta_0 = 13^\circ \text{ N}$ ($k = 2$)	5 yr	234.5 ± 5.2	109.5 ± 1.6
	$\theta_0 = 16^\circ \text{ N}$ ($k = 2$)	5 yr	262.8 ± 10.3	123.5 ± 5.0
	$\theta_0 = 10^\circ \text{ N}$ ($k = 3$)	5 yr	208.6 ± 3.8	99.6 ± 2.7
	$\theta_0 = 13^\circ \text{ N}$ ($k = 3$)	5 yr	246.8 ± 3.4	120.7 ± 4.0
	$\theta_0 = 16^\circ \text{ N}$ ($k = 3$)	5 yr	268.9 ± 6.9	140.2 ± 6.3

Table 5.1: A summary of the HiRAM asymmetric aquaplanet simulations (control runs and zonal wavenumber $k = \{1, 2, 3\}$ perturbation runs), and the resulting mean genesis frequency of the TCs tracked within each simulation. The frequency is the rate of formation equatorward of 30° N for all TCs ($\mathbf{v}_{\text{sfc}} > 15.2 \text{ m s}^{-1}$) and strong TCs ($\mathbf{v}_{\text{sfc}} > 29.5 \text{ m s}^{-1}$), with the mean and standard error computed over the length of the simulation.

The results in Table 5.1 show that substantial changes occur in the genesis frequency with the introduction of zonal variation to the underlying SST. These changes can be visualized in Figure 5.1, where the annual genesis frequency has been plotted against the zonal wavenumber of the SST perturbation. Of immediate interest to note here is the robust result of increased TC genesis frequency across all of the HiRAM aquaplanet simulations when the SST pattern is shifted away from the equator ($\theta_0 = 10^\circ \text{N} \rightarrow 13^\circ \text{N} \rightarrow 16^\circ \text{N}$). Since this result was a focus of the discussion in Section 3.2 it will not be explored much further here. However it is certainly worth noting that this important earlier result holds more generally for the zonally-asymmetric climates investigated in the current chapter.

The second thing to note from Figure 5.1 (and Tab. 5.1) is that the sensitivity of the mean genesis frequency to the latitude of the maximum SST reduces with the introduction of a zonally-varying SST perturbation. In the zonally-symmetric control

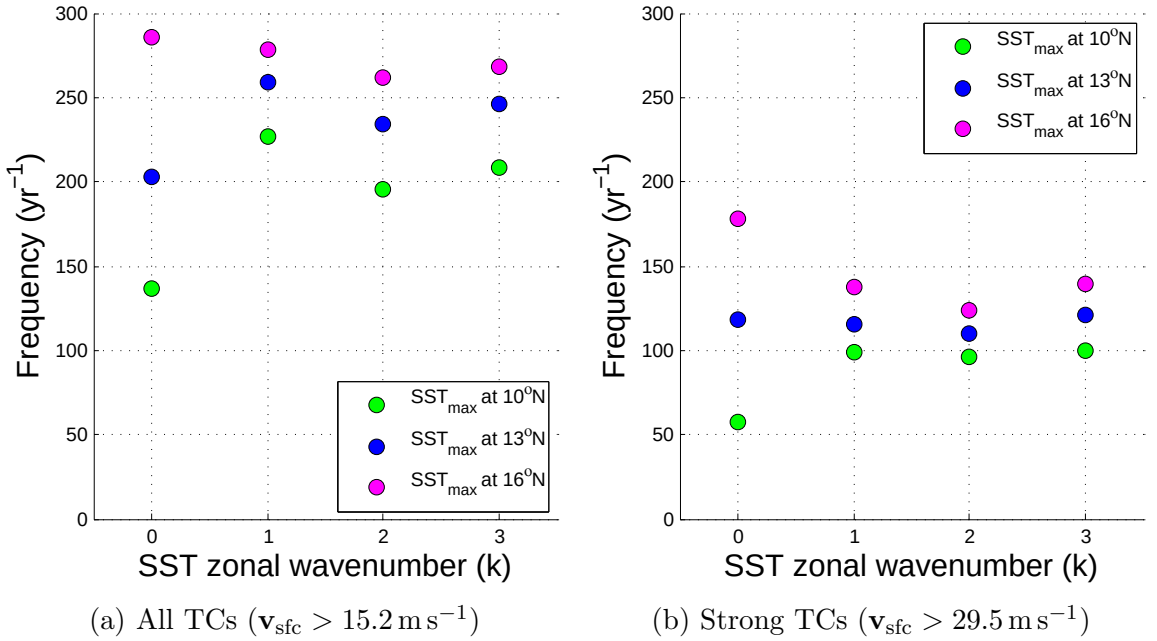


Figure 5.1: The mean genesis frequency in the HiRAM aquaplanet fixed-SST control ($k = 0$) simulations and zonal wavenumber $k = \{1, 2, 3\}$ SST perturbation simulations. Only TCs that formed equatorward of 30°N are included.

runs the genesis frequency (for all TCs) increases by $\sim 50\%$ between the $\theta_0 = 10^\circ \text{ N}$ simulation and the $\theta_0 = 13^\circ \text{ N}$ simulation, and more than doubles between $\theta_0 = 10^\circ \text{ N}$ and $\theta_0 = 16^\circ \text{ N}$. The fractional change in the frequency of hurricane-strength TCs within the control runs is even greater, with the numbers doubling and tripling with the respective shifts in θ_0 . In contrast, the proportional change within the zonally-asymmetric experiments (all TCs, and strong TCs) is much smaller, exhibiting only a $\sim 30\text{-}40\%$ increase when the latitude of maximum SST is moved from 10° N to 16° N in the $k = \{1, 2, 3\}$ experiments.

Thirdly, the results from the different zonally-symmetric control experiments (having $\theta_0 = \{10^\circ, 13^\circ, 16^\circ\}$) span a large range of genesis frequencies ($\sim 135\text{-}286 \text{ TCs yr}^{-1}$), encompassing the smaller range of frequencies realized across all of the zonally-asymmetric experiments ($\sim 195\text{-}277 \text{ TCs yr}^{-1}$). In other words, the impact of the asymmetric perturbation on genesis frequency differs in sign between the three control simulations. Whereas a $k = \{1, 2, 3\}$ SST perturbation significantly increases the global frequency of TC genesis (compared with the zonally-symmetric climate) when the latitude of the maximum SST is set at 10° N , it instead acts to decrease the overall frequency of TCs in the $\theta_0 = 16^\circ \text{ N}$ simulations.

Finally, note that only relatively small changes in the global frequency occur as the SST perturbation wavenumber is increased from $k = 1$ to $k = 2$ and $k = 3$. These changes in total frequency are of order 10% , and smaller than that associated with a 3° shift in the latitude of maximum SST. The changes are also nonmonotonic with perturbation wavenumber, with a minimum occurring in the $k = 2$ simulation for all three sets of experiments ($\theta_0 = \{10^\circ, 13^\circ, 16^\circ \text{ N}\}$). While changes do occur in the geographical distribution of genesis locations (to be discussed next), these results suggest that the total number of TCs forming within the zonally-asymmetric simulations remains relatively insensitive to the number of zonal SST maxima, or the associated changes in the mean magnitude of the zonal SST gradient.

5.2.2 Pattern of TC genesis

We next investigate changes in the geographical pattern of genesis frequency in the simulations forced with zonally-asymmetric SSTs, anticipating differences in the distribution of genesis locations compared with those in the control simulations. The pattern of $SST'(\lambda, \theta)$ that is applied as the lower boundary condition in the nine zonally-asymmetric simulations is plotted in Figure 5.2 (shading) and Figure 5.3 (contours). The black dots overlaid (Fig. 5.2) indicate the genesis locations of all TCs detected in each of the 5-yr simulations. For comparison the genesis locations from each of the three 10-yr control simulations are shown across the top panels.

It is immediately apparent from Figure 5.2 that TCs preferentially form over those regions with an underlying positive SST perturbation, marked by the clustering of black dots over red regions and the relative scarcity of points over blue regions. In order to more clearly depict the clustering, density maps showing the frequency of TC genesis are provided in Figure 5.3. For these plots the longitude dimension (abscissa) has been re-scaled to $\hat{\lambda}$, a normalized distance (in radians) extending the length of one zonal period of the underlying SST perturbation. The fields have been aligned such that $SST'(\hat{\lambda}) \sim \cos(\hat{\lambda})$, having a zonal maximum and minimum in SST' located at $\hat{\lambda} = 0$ and $\hat{\lambda} = \pm\pi$, respectively. Red (positive) and blue (negative) SST' contours show the pattern of underlying perturbation SST, contoured at $\pm 0.3 K$.

The shading within each panel (Fig. 5.3) shows the frequency of TC genesis (yr^{-1}) per grid box area, where each grid box extends 2° in latitude and $\pi/16$ radians in equivalent longitude. For $k = \{1, 2, 3\}$ this zonal box width is equivalent to $\Delta\lambda = \{11.25^\circ, 5.625^\circ, 3.75^\circ\}$, however for $k > 1$ the TC counts over each of the k zonal periods have been accumulated prior to plotting such that the integral over the displayed frequencies yields the total global frequency that is reflected in Fig. 5.1a. Hence these frequency maps can be directly compared between the simulations, each

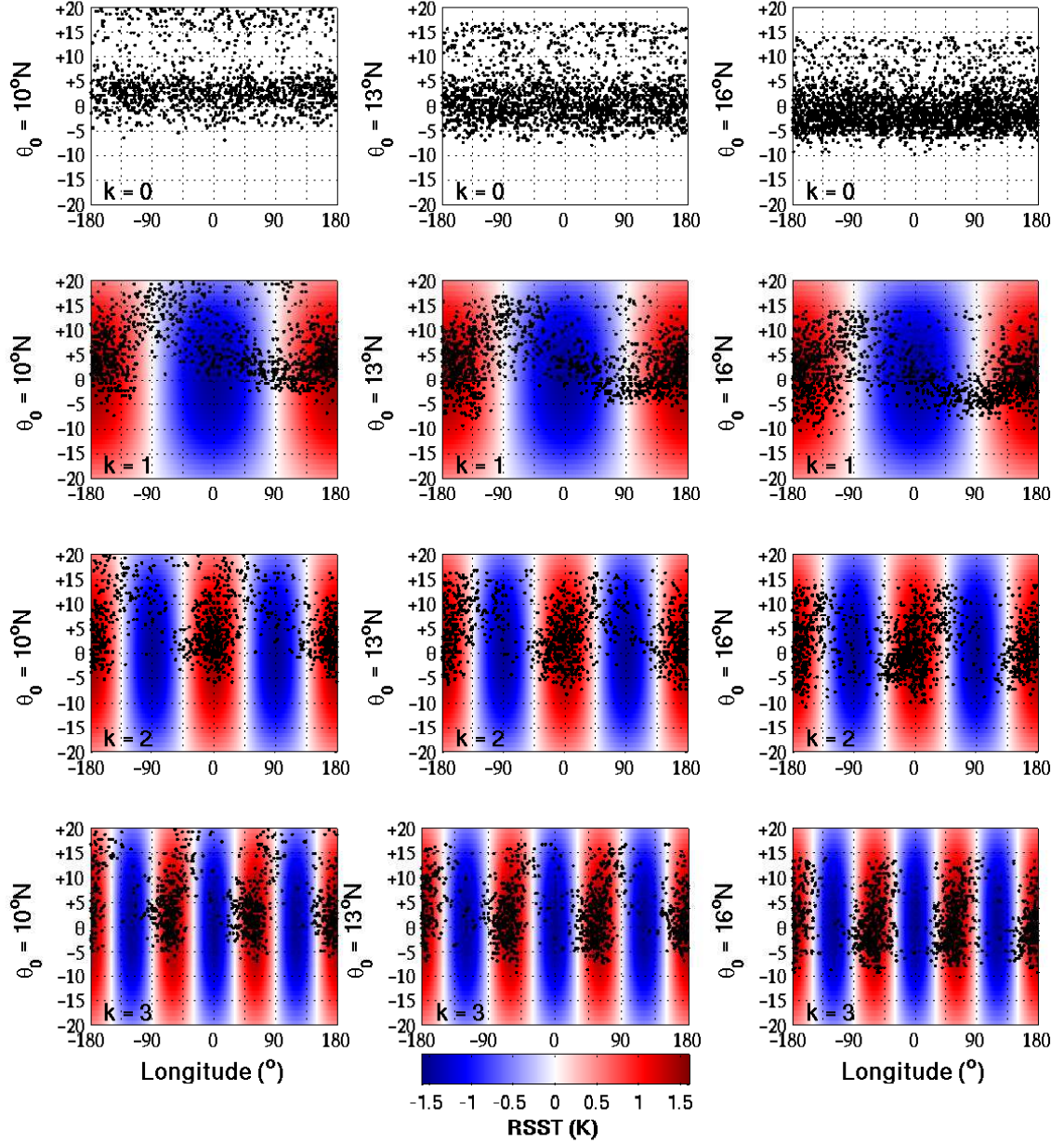


Figure 5.2: The geographical distribution of tropical cyclone genesis locations (black dots) detected in the three zonally-symmetric control simulations and nine zonally-asymmetric simulations. Shading indicates the prescribed temperature perturbation ($\pm 1.5\text{K}$ amplitude) to the control run SSTs for $\theta_0 = 10^\circ\text{N}$ (left column), $\theta_0 = 13^\circ\text{N}$ (center column), and $\theta_0 = 16^\circ\text{N}$ (right column), for zonal perturbation wavenumber $k = \{1, 2, 3\}$. The ordinate displays the meridional direction, with latitude plotted relative to the latitude of the maximum SST. Only TCs that form equatorward of 30°N have been plotted.

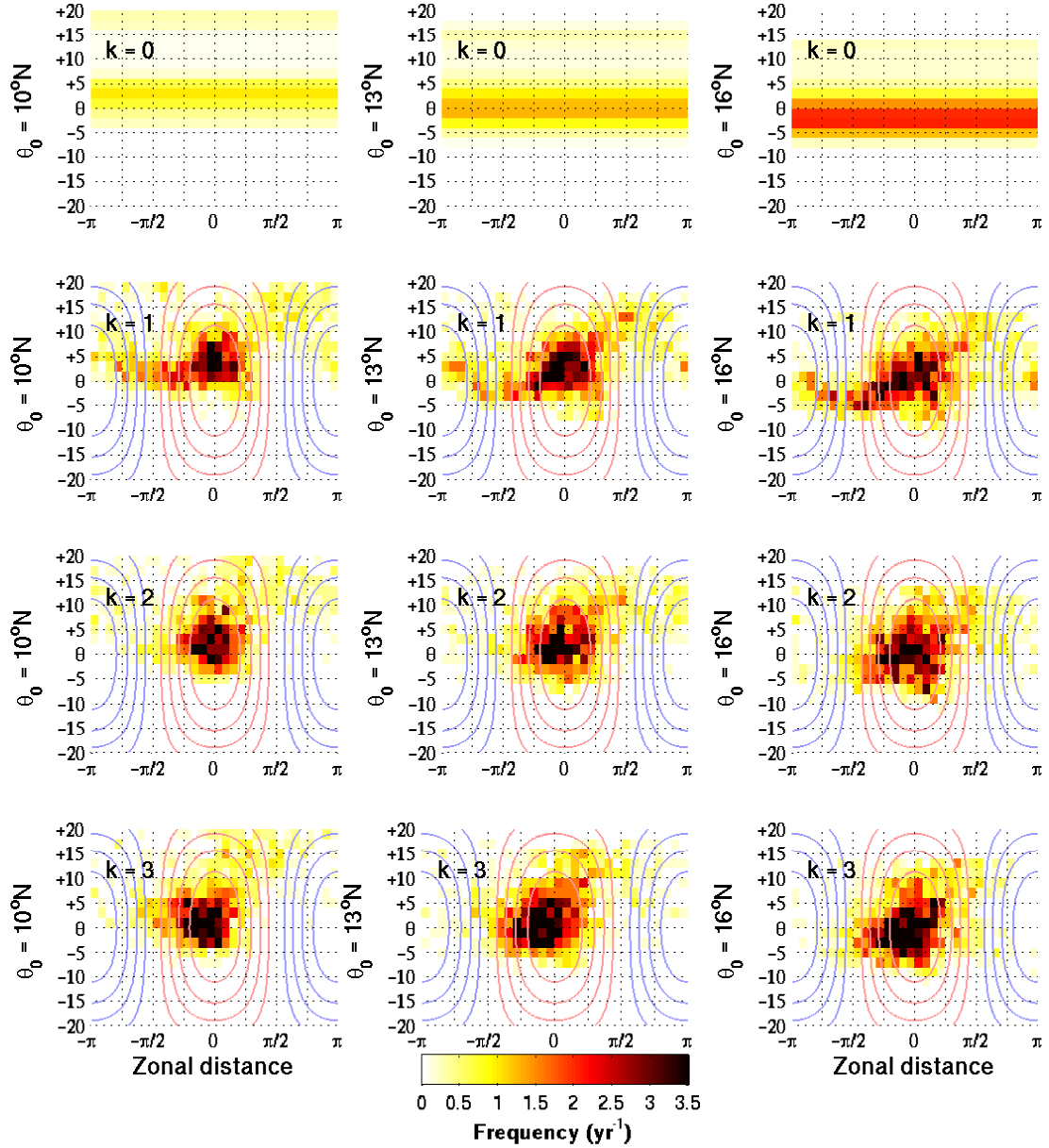


Figure 5.3: The relative density of tropical cyclone genesis locations (shading) detected in the nine zonally asymmetric HiRAM aquaplanet simulations, along with the three control runs. Thin red (positive) and blue (negative) contours show the pattern of underlying perturbation SST, contoured at $\pm 0.3\text{ K}$. The ordinate displays the meridional direction, with latitude plotted relative to the latitude of the maximum SST. The abscissa is a normalized distance $\hat{\lambda}$ (in radians), aligned relative to the underlying SST perturbation pattern with the zonal maxima and minima (in SST) located at $\hat{\lambda} = 0$ and $\hat{\lambda} = \pm\pi$, respectively. For the zonally-symmetric control runs ($k = 0$ plots) the zonal mean frequency has also been normalized to enable comparison of the relative occurrence density.

having the equivalent units of TC genesis counts per year per $k \times (2^\circ \text{lat} \times \frac{11.25^\circ}{k} \text{lon})$ boxes (per $\sim 270,000 \text{ km}^2$).

The patterns of genesis density in Figure 5.3 provide some further insights. Firstly, compared with the zonally-uniform control simulations, genesis within the zonally-asymmetric SST experiments is itself zonally-asymmetric, having a zonal peak in genesis close to the maximum in perturbation SST. The genesis is largely confined within the region of positive SST anomaly. The pattern in the $k = 1$ simulations extends westward into the region of the cool anomaly, along the same latitudinal band where the majority of TCs occur in the corresponding $k = 0$ control runs. The $k = 1$ genesis density pattern also extends into the subtropics through the northeast quadrant of the positive SST anomaly, with a significant number of storms occurring in the region around $\hat{\lambda} \approx \pi/2$ and $\theta \approx \theta_0 + 10^\circ$ (where $SST' \approx 0$). The asymmetrical elongation that is evident across the $k = 1$ simulations reduces substantially as the wavenumber of the SST perturbation increases. For the $k = 2$ and $k = 3$ simulations the pattern contracts and becomes more circularly symmetric, with most TCs occurring within the higher values of perturbation SST.

5.2.3 RSST and TC Occurrence Density

To analyze the concentration of TC genesis region within the anomalously warm regions we will use the distribution of relative SST (RSST) values coincident with the TC genesis locations. Here the RSST is defined simply by $SST'(\lambda, \theta)$ (Eq. 5.1), the prescribed perturbation SST. By construction, the total areal distribution of an RSST interval within the $\theta_0 \pm 30^\circ$ tropical band remains unchanged between the simulations for $k = \{1, 2, 3\}$, given the amplitude $(SST'(\theta)|_{\text{sym}})$ is the same. The relative area of the broad tropical region $\theta_0 \pm 30^\circ$ that has RSST values in excess of a certain threshold value of RSST is plotted in Figure 5.4. By definition, the application of a sinusoidal perturbation in SST (along λ) yields a relative area of 0.5 for a threshold

value of RSST = 0 K, meaning half of the perturbed region has RSST > 0 K. As the threshold value of RSST increases towards the maximum RSST, the relative tropical area in excess of this value reduces to zero. The change in area is not dependent on k , so the curves for the $k = \{1, 2, 3\}$ simulations are identical.

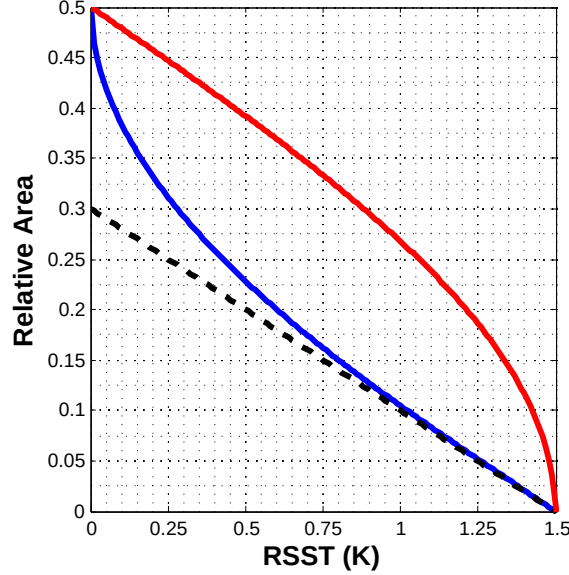


Figure 5.4: The relative area of the broad tropical region ($\theta_0 \pm 30^\circ$) in excess of a certain value of RSST, plotted as a function of RSST. See text for further details.

The red curve in Figure 5.4 shows the relative area of the rectangular region(s) bounded (east/west) by the longitudes corresponding to the threshold values of RSST and (north/south) by the latitudes $\theta_0 \pm 30^\circ$, and is thus simply proportional to $\cos^{-1}(\text{RSST})$. However, due to the additional modulation of the SST perturbation with latitude (Eq. 2.2), the actual relative area in excess of a particular RSST decreases more rapidly, as shown by the blue curve. This curve shows (for example) that a quarter of the tropical region has a relative SST in excess of $\sim +0.4$ K, and approximately 10% of the area has RSST > +1 K. The accompanying dashed black line shows a linear approximation for the change in relative area for large RSST values. One can see that for RSST values greater than $\sim +1$ K, a 10% increase in RSST threshold yields a 20% reduction in area.

We next compute the cumulative distribution of RSST values sampled at every TC genesis location for each of the simulations. The resulting curves plotted in Figure 5.5 show the proportion of TCs that form over the region (or regions, for $k > 1$) with a RSST in excess of a certain value (abscissa). Thin lines show the individual simulations $\theta_0 = \{10^\circ, 13^\circ, 16^\circ \text{ N}\}$, but for clarity we will focus here on the combined results for the $k = 1$, $k = 2$ and $k = 3$ sets of experiments (thick lines). Figure 5.5 shows that the proportion of TCs that form within any specified positive contour of RSST increases as k increases between experiments.

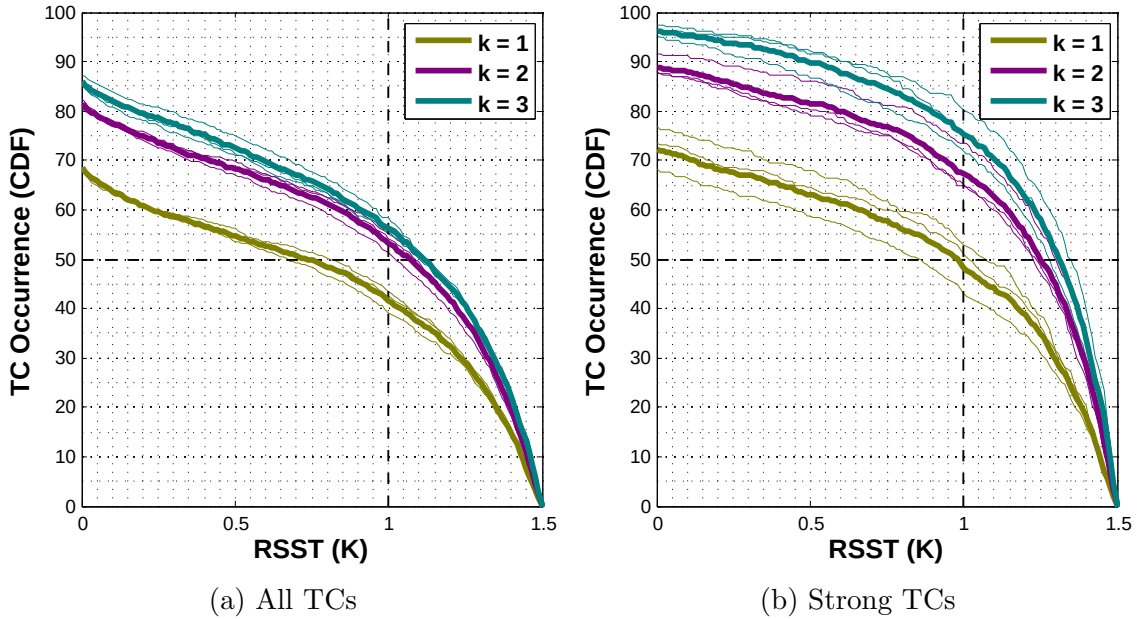


Figure 5.5: A cumulative distribution showing the proportion of TCs (%) that form over the region(s) having a RSST (K) in excess of a certain value, for the different zonally-asymmetric SST perturbation simulations. Thin lines show the three different $\theta_0 = \{10^\circ, 13^\circ, 16^\circ \text{ N}\}$ experiments, and thick lines display the average distribution across those three simulations.

We could consider, for example, the proportion of TCs within the contour of $\text{RSST} > +1 \text{ K}$ (dashed vertical line). As k increases from one to three, the proportion of all TCs that form within that region increases from $\sim 42\%$ to $\sim 57\%$ (Fig. 5.5a). The subset of stronger hurricane-strength TCs (Fig. 5.5b) preferentially form over higher RSSTs, as shown by the higher percentages across all RSST values. Furthermore, as

k increases the relative proportion of strong TCs to all TCs increases. As k increases from one to three, the proportion of strong TCs that form within the $+1$ K region increases from $\sim 50\%$ to $\sim 76\%$.

Alternatively we can look at changes to the size of the region containing the same proportion of TCs. For example, we can read off the RSST contours containing 50% of the TCs (Fig. 5.5, dashed horizontal line) for the different simulations, and see that increasing k moves the distribution towards warmer values of RSST. This contraction is visualized in Figure 5.6, where the RSST along the θ_0 latitude circle is shown by the red-blue curves. The colored horizontal lines, corresponding to the $k = \{1, 2, 3\}$ simulations at $\theta_0 = 16^\circ$ N, show the median RSST (solid lines) from the distribution of RSST values sampled at each of the TC genesis locations, along with the 25th and 75th percentiles of RSST (dashed lines). From Figure 5.6 we see that 75% of TCs in the $k = 2$ simulation have their point of genesis located over RSST values in excess of $\sim +0.2$ K; half of the total number of TCs form over a region with $\text{RSST} \gtrsim +1$ K, and half of these occur over $\text{RSST} \gtrsim +1.3$ K.

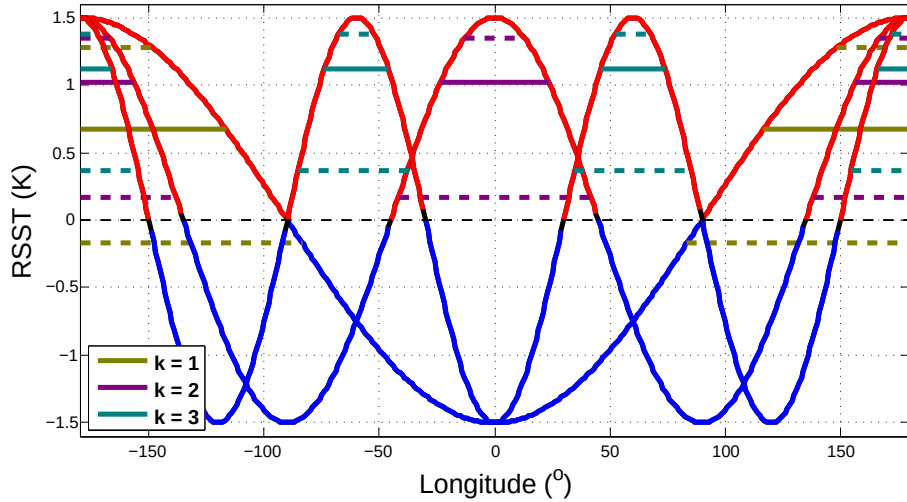


Figure 5.6: The median RSST (solid horizontal lines) from the distribution of RSST values sampled at each of the TC genesis locations in the three zonally-asymmetric SST perturbation simulations ($k = \{1, 2, 3\}$) at $\theta_0 = 16^\circ$ N, along with the 25th and 75th percentiles of RSST (dashed lines).

From the relationship between the RSST threshold and relative area of the tropics (Fig. 5.4), and of the RSST threshold with the proportion of TC occurrence (Fig. 5.5), we are able to compute the proportion of TCs as a function of the relative area of that region, shown in Figure 5.7. For a more quantitative measure, the absolute cumulative frequency of TCs (yr^{-1}) has been divided by the absolute area of the enclosed region ($\times 10^6 \text{ km}^2$) to produce a TC occurrence density, shown in Figure 5.8.

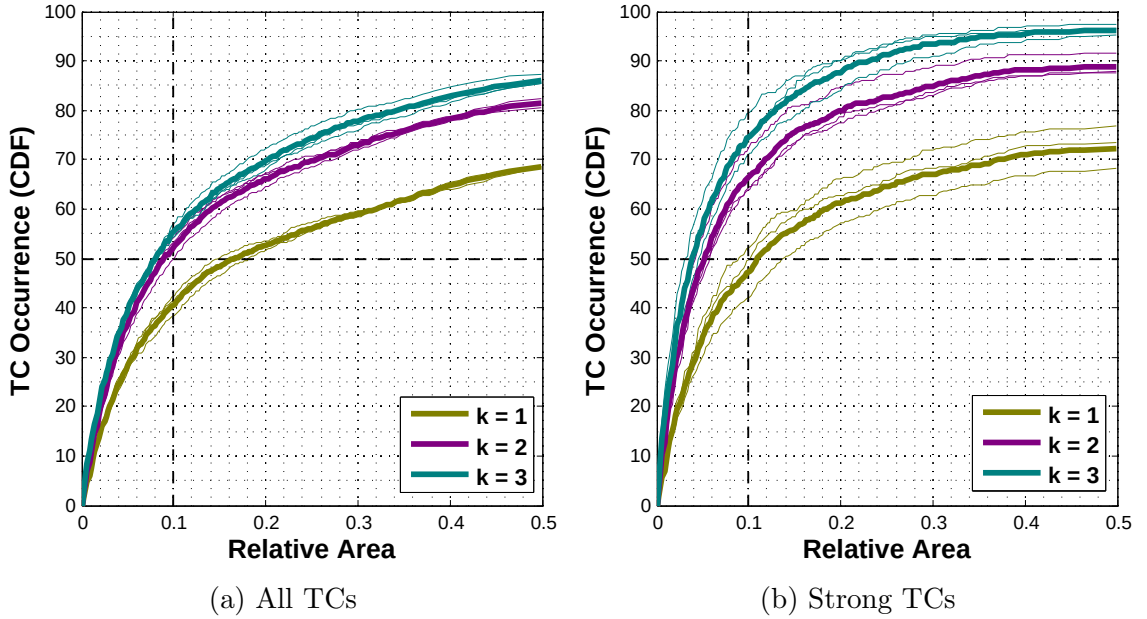


Figure 5.7: A cumulative distribution showing the proportion of TCs (%) that form over the warmest region(s) (defined by contours of RSST), as a function of the area of that region relative to the $\theta_0 \pm 30^\circ$ tropical band, for the different zonally-asymmetric SST perturbation simulations. The relative area increases with decreasing RSST threshold (Fig. 5.4). Thin lines show the three different $\theta_0 = \{10^\circ, 13^\circ, 16^\circ \text{ N}\}$ experiments, and thick lines display the average distribution across those three simulations.

Figures 5.7 and 5.8 show how the relative proportion of TCs and the implied mean genesis density change as the wavenumber of the SST perturbation patterns increases. As k increases the relative area of genesis (considering all TCs) contracts (Fig. 5.7a) and the absolute TC occurrence density increases within those regions (Fig. 5.8a). The impact of increasing k on the subset of stronger TCs is even greater, with a marked decrease in the area containing the majority of genesis locations (Fig. 5.7b)

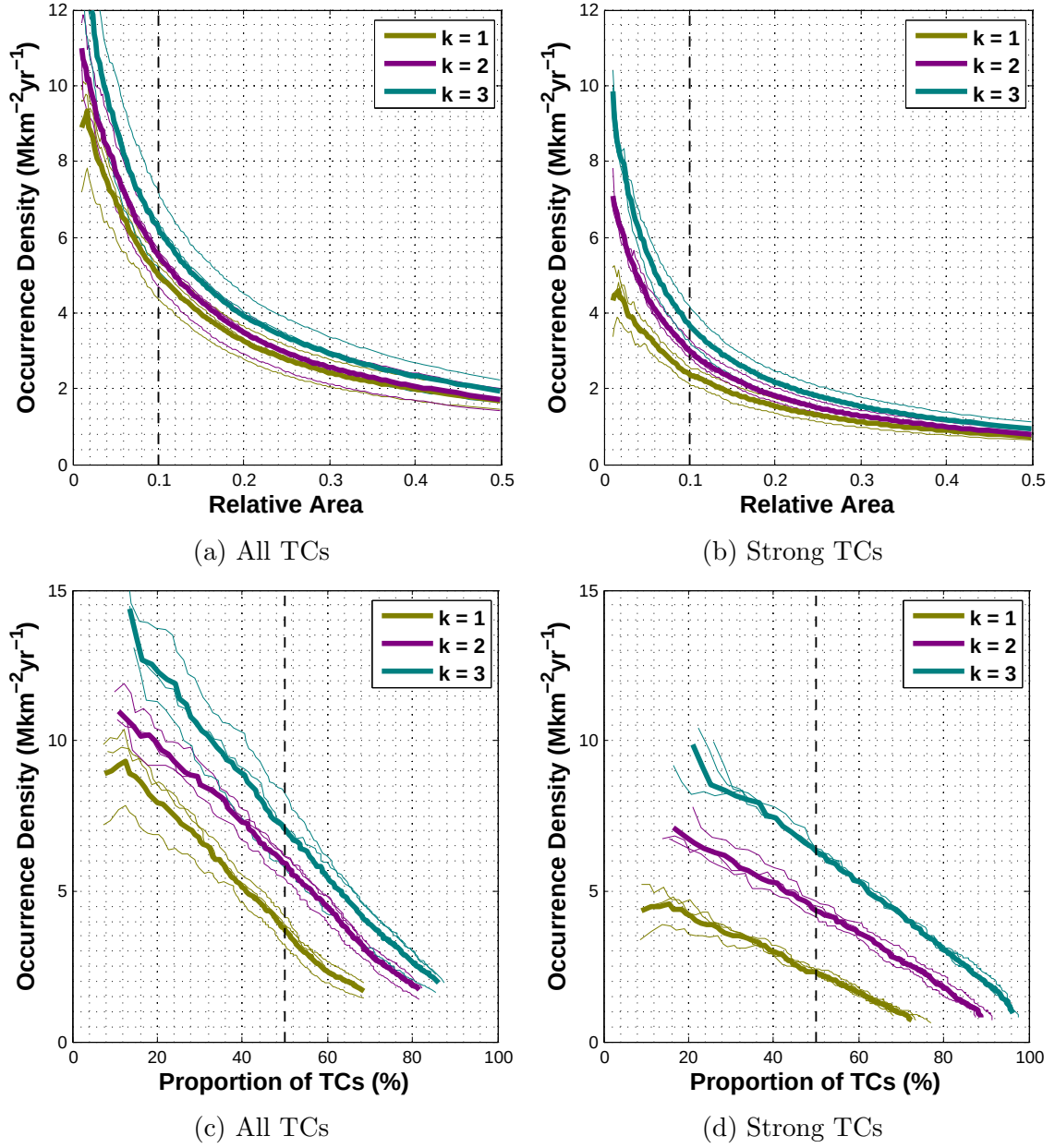


Figure 5.8: The occurrence density of TCs (per 10^6 km^2 per year) that form over the warmest region(s) (defined by contours of RSST) for the different zonally-asymmetric SST perturbation simulations. Plotted against the area of that region relative to the $\theta_0 \pm 30^\circ$ tropical band (a and b), and the proportion of the total number of TCs (c and d). Thin lines show the three different $\theta_0 = \{10^\circ, 13^\circ, 16^\circ \text{ N}\}$ experiments, and thick lines display the average occurrence density across those three simulations.

and a greater proportional increase in the occurrence density of strong TCs (Fig. 5.8b) in the regions of high RSST.

From the curves in Figure 5.8c and 5.8d, we next compute the TC genesis occurrence density over the region(s) enclosed by the RSST contour containing 50% of TC genesis locations (dashed vertical lines) for that particular simulation, and plot it against the perturbation wavenumber k , as shown in Figure 5.9. It is difficult to compute an equivalent occurrence density for the $k = 0$ control simulations, since $\text{RSST} = 0$ everywhere. For the zonally-symmetric ($k = 0$) control runs in Figure 5.9, the occurrence density plotted is simply the frequency of genesis divided by the area of the entire $\theta_0 \pm 30^\circ$ latitude band. We must note this only provides a lower-bound on the average occurrence density, but the values are nonetheless included here for completeness.

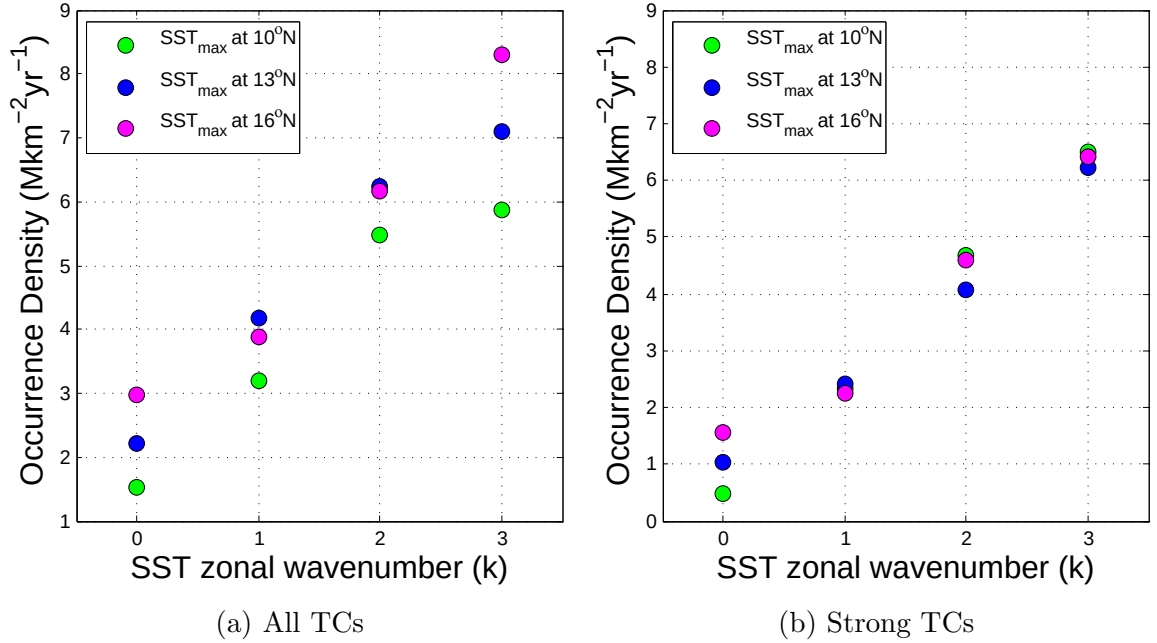


Figure 5.9: TC genesis occurrence density in the HiRAM simulations as a function of SST perturbation wavenumber k , computed over the $\theta_0 \pm 30^\circ$ latitude band. For the $k \geq 1$ simulations the occurrence density is computed over the region(s) enclosed by the constant RSST contour containing 50% of TC genesis locations. Units displayed are in TCs per 10^6 km^2 per year.

Despite the relatively small changes in global TC frequency between the $k = \{1, 2, 3\}$ zonally-asymmetric SST perturbation simulations (Fig. 5.1), in Figure 5.9 we see that there are in fact large changes in the density of TC occurrence within

the preferred regions of genesis. Take for example the set of simulations conducted at $\theta_0 = 16^\circ \text{ N}$, where, in the case of the $k = 1$ simulation, half of the TCs form in the region with RSST values in excess of $\sim +0.7 \text{ K}$ (Fig. 5.6), which represents $\sim 15\text{-}20\%$ of the tropical area relative to the $\theta_0 \pm 30^\circ$ latitudinal band (Fig. 5.4). Within this particular region the frequency of TC genesis occurrence is $\sim 4 \text{ Mkm}^{-2}\text{yr}^{-1}$ (Fig. 5.9a).

As k increases the areal region containing half of the TCs decreases to $\sim 10\%$ and $\sim 7\%$ of the tropics in the $k = 2$ and $k = 3$ simulations, respectively (Fig. 5.6, 5.4). Along with this contraction of the primary genesis region itself, the frequency of genesis increases to $\sim 6 \text{ Mkm}^{-2}\text{yr}^{-1}$ ($k = 2$) and $\sim 8 \text{ Mkm}^{-2}\text{yr}^{-1}$ ($k = 3$) within those regions (Fig. 5.9a). The same occurs with the subset of hurricane-strength TCs, with the frequency of occurrence increasing quasi-linearly between simulations with increasing k (Fig. 5.9b). Note that the region containing 50% of the strong TCs (not shown) is smaller (having a higher average RSST) than the corresponding region for all TCs, hence a quantitative comparison between the two plots (Fig. 5.9a, 5.9b) must bear this in mind.

The occurrence densities corresponding to the different zonally-asymmetric simulations are plotted again in Figure 5.10, but in this figure they have been computed over the same $\text{RSST} > +1 \text{ K}$ region, which has an area of $\sim 26 \text{ Mkm}^2$ in each of the $k = \{1, 2, 3\}$ simulations. Fixing the area means that different proportions of the developing TCs are included in the calculation of genesis occurrence density. However, it allows a direct absolute comparison of the actual TC density within a fixed region, neglecting the associated contribution from the different sizes of the regions.

The markers in Figure 5.10a show that the frequency of genesis within the warmest 10% of the tropics ($\text{RSST} > +1 \text{ K}$) increases as the wavenumber of the SST perturbation increases. Moving from the $k = 1$ SST pattern to $k = 3$ increases the frequency of genesis in this region by $\sim 1\text{-}2 \text{ Mkm}^{-2}\text{yr}^{-1}$ within this suite of experiments.

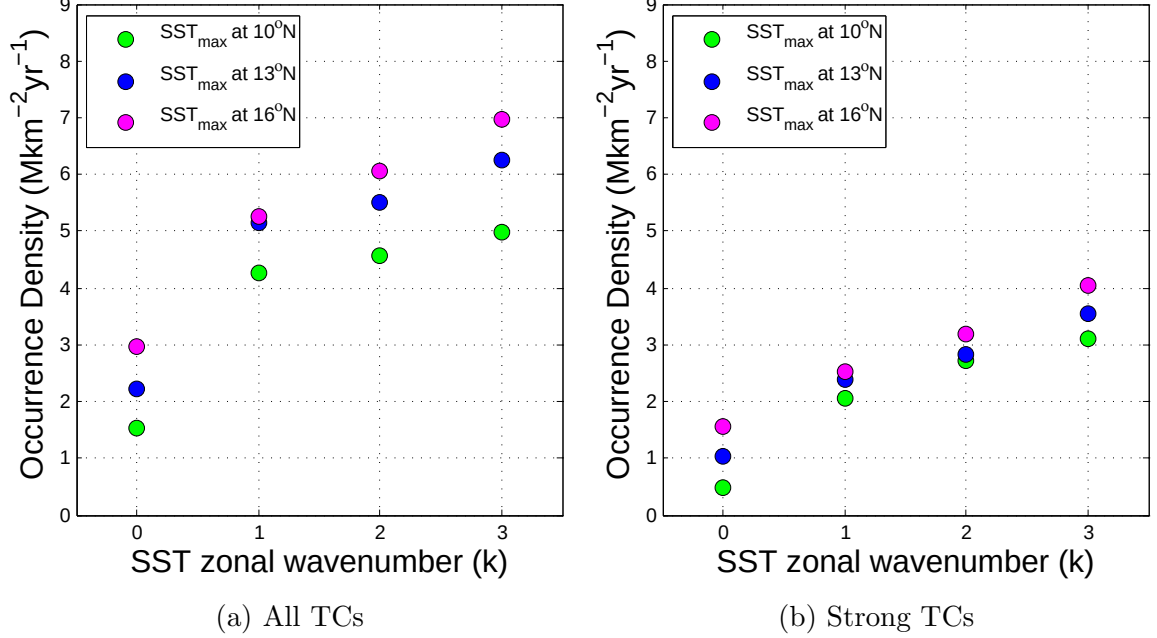


Figure 5.10: TC genesis occurrence density in the HiRAM simulations as a function of SST perturbation wavenumber k , computed over the $\theta_0 \pm 30^\circ$ latitude band. For the $k \geq 1$ simulations the occurrence density is computed over the region(s) enclosed by the 1 K RSST contour. Units displayed are in TCs per 10^6 km^2 per year.

This change appears to account for roughly half of the total change in the frequency of occurrence density found when holding the proportion of TCs fixed (Fig. 5.9a). A similar result is observed for the stronger subset of TCs displayed in Figure 5.10b. The increasing frequency of occurrence (holding the area fixed) as k increases between simulations (Fig. 5.10) shows that the increasing density previously discussed (Fig. 5.9) can be attributed both in part to the reduction of the actual genesis area, and also in part to a real absolute increase in TC genesis frequency within the warmest tropical regions.

In summary, these results suggest that the proclivity of TCs to form in regions of higher RSST depends not only on the underlying values of RSST, but also on the regional zonal gradient of RSST (or SST). As $|\overline{\nabla_\lambda SST}|$ increases (linearly with k), TC formation is *pushed* towards regions of higher RSST, increasing the TC occurrence density within these shrinking regions.

5.2.4 Large-scale environmental fields

Genesis-weighted fields

Estimates of the genesis-weighted environmental fields are shown in Figure 5.11. These distributions give an indication of the spread in the values of the time-mean climate coincident with the location of the TCs that form within each simulation. Firstly a simple mean is computed over a sample of 10 TC genesis locations (randomly chosen with replacement). Further bootstrap re-samples ($N = 10^5$) of this mean statistic are computed to form the distribution which is being displayed in Figure 5.11. The distributions have been normalized (unit area under the curve) to enable comparison as estimates of the probability density function of the genesis-weighted mean. The mode of each of these quasi-normal distributions indicates the genesis-weighted mean over the entire simulation, and the relative spread in the statistic indicates the relative variation in the environmental field at the genesis locations.

Figure 5.11a shows the genesis-weighted relative sea surface temperature $\langle \text{RSST} \rangle_G$, where the RSST has been previously defined as the local perturbation from the zonal-mean SST. Since by definition the RSST field in the zonally-symmetric control experiments is identically zero it has not been plotted. Values of RSST in the asymmetric experiments, $k = \{1, 2, 3\}$, are bounded by the magnitude of the perturbation (± 1.5 K in all experiments), and as expected, is strongly positive, showing that TCs within these simulations preferentially form over regions of SST that are warmer than the zonal mean.

As discussed in the previous section, these results (Fig. 5.11a) also show that $\langle \text{RSST} \rangle_G$ increases as k increases. For the $k = 0$ (zonally-symmetric) simulations the $\langle \text{RSST} \rangle_G = 0$ K, increasing to $\langle \text{RSST} \rangle_G \approx 0.7$ K for $k = 1$, $\langle \text{RSST} \rangle_G \approx 0.9$ K for $k = 2$, and $\langle \text{RSST} \rangle_G \approx 1.1$ K for the $k = 3$ simulation. This shift is not due to a change in the absolute geographical area that exceeds some particular RSST

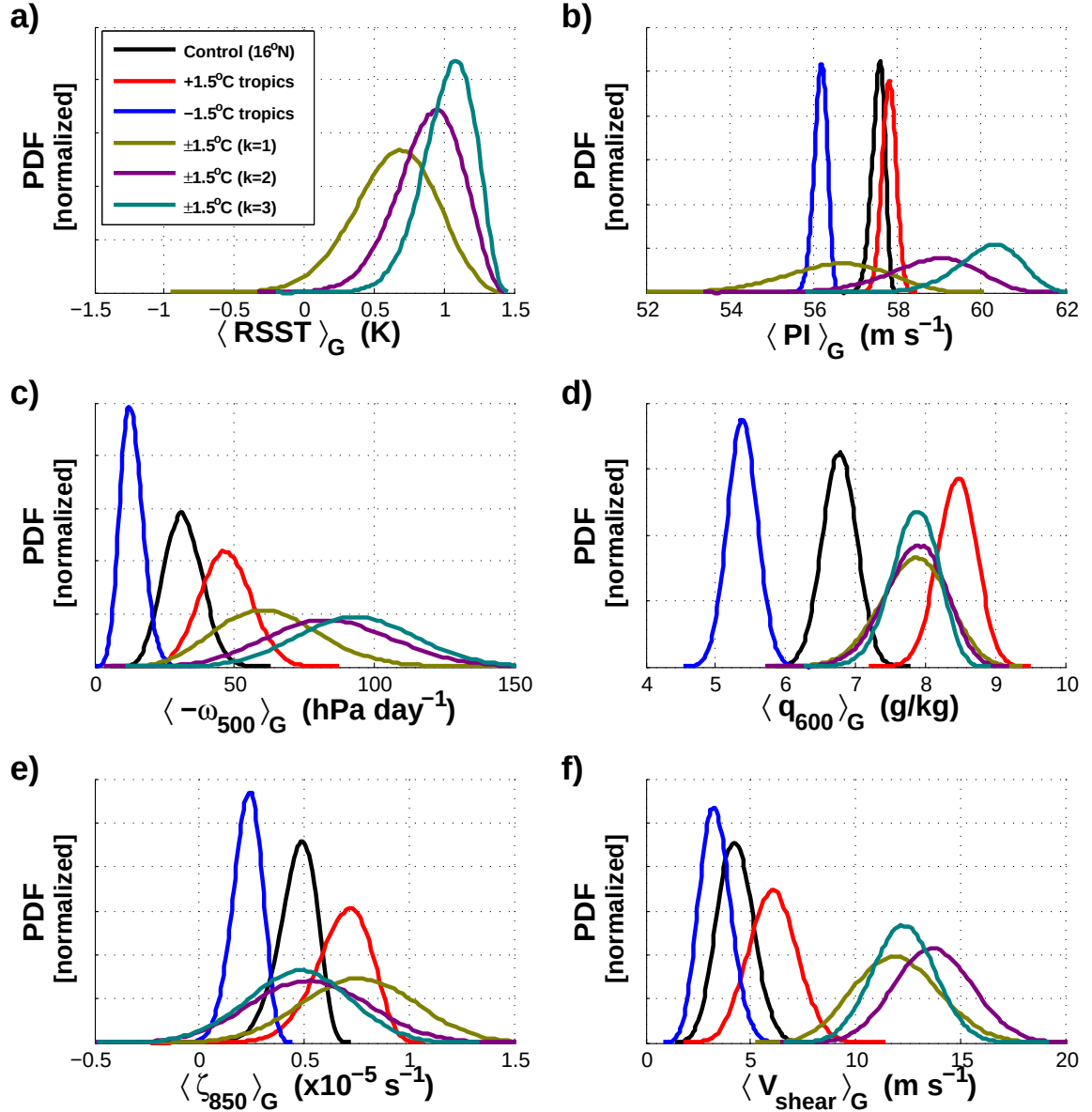


Figure 5.11: The genesis-weighted mean probability density functions for various environmental fields, estimated by bootstrap sampling ($N = 10^5$) of the statistic, being the mean value of the time-mean environmental field sampled at the genesis locations of 10 of the simulated TCs (randomly chosen with replacement). Plots show the relative sea surface temperature (a, K), potential intensity (b, m s^{-1}), pressure velocity at 500 hPa (c, hPa day^{-1}), specific humidity at 600 hPa (d, g/kg), relative vorticity at 850 hPa (e, $\times 10^{-5} \text{ s}^{-1}$), and vertical shear (200-850 hPa) of the horizontal wind (f, m s^{-1}).

threshold, since by construction (integer periods of a cosine perturbation) such an area is identical between the experiments (with different k). Rather, this change indicates a relative contraction of the total region of genesis as the TCs preferentially form over the warmer SSTs. This result could be similarly displayed in a plot of the absolute $\langle \text{RSST} \rangle_{\text{G}}$ (not shown here).

The genesis-weighted potential intensity $\langle \text{PI} \rangle_{\text{G}}$ is shown in Figure 5.11b. The sharply-peaked narrow distributions of the zonally-symmetric experiments reflect the relatively small range of PI values found in the genesis region on the northern flank of the ITCZ. A small increase in $\langle \text{PI} \rangle_{\text{G}}$ occurs for the tropical warming simulation, compared to a marked decrease for the tropical cooling scenario. The PDF of the $k = 1$ asymmetric simulation is much broader, with TC formation coincident with a wide range of climatological PI values. The $k = 1$ genesis-weighted mean falls between the respective means from the cool-warm tropical perturbation simulations. However, as the perturbation wavenumber increases the $\langle \text{PI} \rangle_{\text{G}}$ distributions shift to higher intensities, consistent with the changes already noted in the surface temperature. The genesis-weighted PI is higher in the $k = \{2, 3\}$ simulations than in the tropical warming experiment.

The genesis-weighted 500-hPa vertical motion and 600-hPa specific humidity are displayed in Figure 5.11c and 5.11d, respectively. Compared with the control, the zonally-symmetric simulations show stronger ascent and moistening of the genesis region in the case of tropical warming, and weaker ascent and drying associated with tropical cooling. The mean vertical motion over the genesis region $\langle \omega_{500} \rangle_{\text{G}}$ is stronger with the introduction of the asymmetric SST perturbations, and, similar to $\langle \text{RSST} \rangle_{\text{G}}$ and $\langle \text{PI} \rangle_{\text{G}}$, increases as k increases. The distribution of $\langle q_{600} \rangle_{\text{G}}$ for the asymmetric simulations lies in between that of the control and tropical warming simulations. Rather than a significant change in the mean value, the impact of increasing k is realized in a narrowing of the distribution.

Figure 5.11e shows the genesis-weighted 850-hPa relative vorticity. The positive horizontal shear of the zonal flow on the northern flank of the ITCZ increases (decreases) with a sharpening (flattening) of the meridional SST gradient in these simulations. The simulation with the highest $\langle \zeta_{850} \rangle_G$ is the one with the $k = 1$ asymmetric SST pattern. The $k = \{2, 3\}$ simulations have a reduced $\langle \zeta_{850} \rangle_G$ which is similar to the zonally-symmetric control simulation, albeit with a larger spread in the distribution of the estimate.

The genesis-weighted vertical wind shear is shown in the final panel, Figure 5.11f. TCs form in relatively low-shear environments ($< 10 \text{ m s}^{-1}$) in the zonally-symmetric simulations, with the lowest values of $\langle V_{sh} \rangle_G$ ($< 5 \text{ m s}^{-1}$) found in the tropical cooling simulation. In stark contrast, TCs experience significant vertical wind shear at their genesis location when a zonally-asymmetric SST anomaly is introduced, with most disturbances forming in regions of vertical wind shear greater than $\sim 10 \text{ m s}^{-1}$. As k increases from 1 to 3 the mean estimate of $\langle V_{sh} \rangle_G$ increases from $\sim 12 \text{ m s}^{-1}$ to $\sim 14 \text{ m s}^{-1}$.

Mean fields over regions exceeding a certain RSST

An alternative approach is now taken in order to determine the mean value of the large-scale atmospheric fields that influence genesis, motivated by the discussion of TC occurrence density in Section 5.2.2 and its relationship with RSST. We first define $\langle X \rangle_{\text{RSST} > 1}$ as the spatial mean of the time-mean field of X , calculated over the geographical region with $\text{RSST} > +1 \text{ K}$. Recall that this region ($\sim 10\%$ of the $\theta_0 \pm 30^\circ$ tropical band) has the same geographical area across all of the zonally-asymmetric perturbation simulations, irrespective of k . Hence $\langle \text{RSST} \rangle_{\text{RSST} > 1} \approx 1.22 \text{ K}$ across all simulations ($k = \{1, 2, 3\}$). Next we define $\langle X \rangle_{50\%}$ as the spatial mean of the time-mean field of X , calculated over the area of the RSST contour containing 50% of TC genesis locations within each simulation. Hence this measure focuses on the warm

region containing the majority of TCs, and contracts with increasing k as previously shown (Sec. 5.2.2).

The density of TC occurrence for each simulation is plotted against these two new measures of genesis-weighted $\langle \text{RSST} \rangle$ in Figure 5.12, where one can see that the occurrence density of TCs increases as the mean RSST (over the contracting region of genesis) increases. The increasing frequency of storms in this region is consistent with the increasing mid-level 500-hPa vertical velocity $\langle -\omega_{500} \rangle$ seen in Figure 5.13, and the related precipitation rate $\langle \text{Precip} \rangle$ in Figure 5.14.

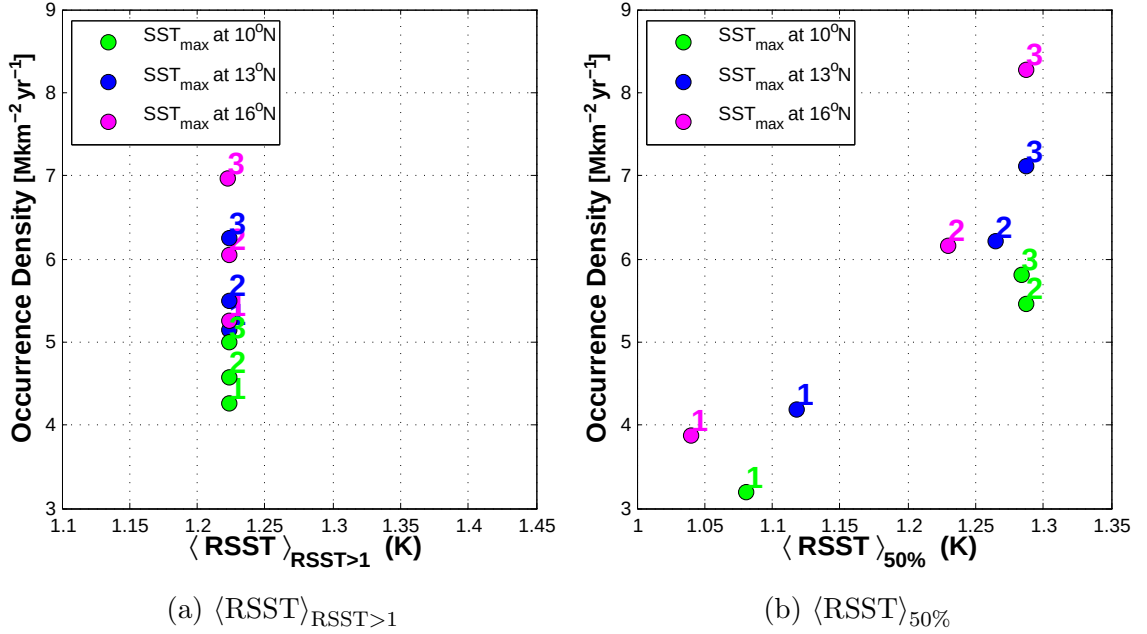


Figure 5.12: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean relative SST, computed over the region(s): a) where $\text{RSST} > +1 \text{ K}$, and b) enclosed by the RSST contour containing 50% of TC genesis locations within that simulation. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

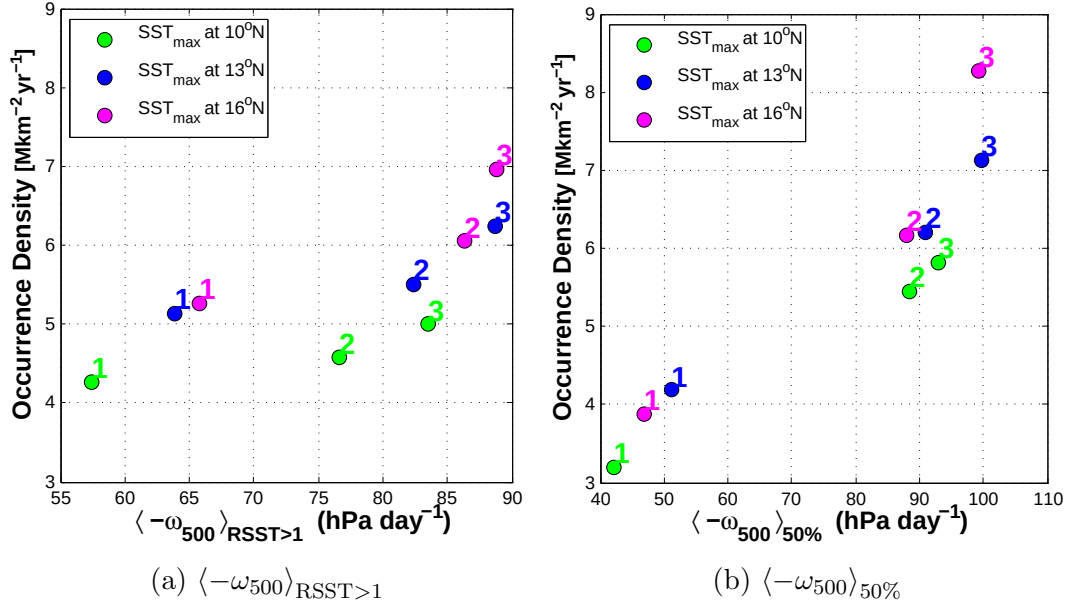


Figure 5.13: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean vertical motion at 500 hPa (hPa day^{-1}), computed over the region(s): a) where $\text{RSST} > +1 \text{ K}$, and b) enclosed by the RSST contour containing 50% of TC genesis locations within that simulation. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

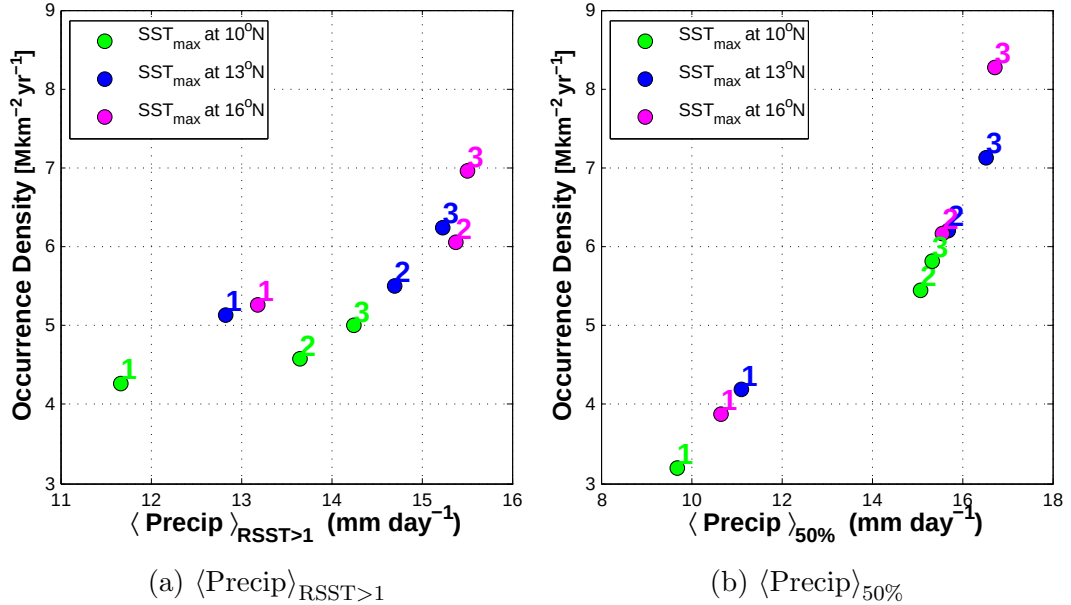


Figure 5.14: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean precipitation rate (mm day^{-1}), computed over the region(s): a) where $\text{RSST} > +1 \text{ K}$, and b) enclosed by the RSST contour containing 50% of TC genesis locations within that simulation. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

5.2.5 Genesis Potential Index

The Genesis Potential Index (GPI) was introduced and developed by Emanuel and Nolan (2004), and is defined as

$$\text{GPI} = |10^5 \eta|^{3/2} \left(\frac{\text{PI}}{70} \right)^3 \left(\frac{\text{RH}}{50} \right)^3 (1 + 0.1V)^{-2} , \quad (5.2)$$

where η is the absolute vorticity at 850 hPa (s^{-1}), PI is the Potential Intensity (m s^{-1}), RH is the relative humidity at 600 hPa (%), and V is the vertical shear (200-850 hPa) of the horizontal wind (m s^{-1}). Hence the GPI incorporates measures of four environmental ingredients that are important for TC formation: 1) sufficient vorticity, 2) thermodynamic potential, 3) abundant moisture, and 4) limited wind shear. In the formulation of GPI each of these factors multiply to give a final estimate of the relative potential for TC genesis.

For each of the first three terms in Eq. 5.2, the absolute impact on GPI (positive or negative) depends on an empirical normalizing factor. The absolute vorticity ($\eta_{850} = \zeta_{850} + f$) incorporates the local relative vorticity (ζ_{850}) and distance from the equator ($f \sim \sin \theta$). The impact on GPI is positive when $\eta_{850} > 10^{-5}$ and negative when $\eta_{850} < 10^{-5}$. Similarly, a Potential Intensity in excess of 70 m s^{-1} makes a positive contribution to GPI, but reduces it when $\text{PI} < 70 \text{ m s}^{-1}$. Thirdly, GPI is increased wherever the environmental relative humidity is greater than 50 %, but reduced where $\text{RH}_{600} < 50 \%$. For the fourth and final term we note that any non-zero value of vertical wind shear acts to reduce the genesis potential.

Since the absolute magnitude of GPI is arbitrary and not very useful quantitatively without reference to a particular observed climate or model simulation, for the following analysis we will focus on the *pattern* of mean GPI, and the *change* in GPI associated with the *change* in the four environmental fields that compose it. The pattern and qualitative scale of the changes can then be compared with the pattern

and magnitude of the change in the genesis frequency of the resulting TCs that are simulated.

We commence by analyzing the GPI changes in the zonally-asymmetric $k = 1$ SST perturbation simulation compared with the corresponding control simulation in order to investigate the impact of a regional perturbation to the zonally-symmetric climate. Following this we analyze the GPI changes in the $k = 3$ simulation compared with the $k = 1$ simulation in order to investigate the impact of increasing k on the pattern of genesis.

GPI pattern for the $k = 1$ perturbation

The mean GPI for the $k = 1$ SST perturbation simulation (with $\theta_0 = 13^\circ \text{ N}$) is shown by the thick black contours in Figure 5.15a, and light gray contours indicate the underlying SST field. The shading shows the actual density of genesis occurrence, which was also plotted earlier in Figure 5.3. The meridional-mean (computed over the $\theta_0 \pm 10^\circ$ latitude band) and zonal-mean profile of GPI is shown above and to the right of each panel, respectively, along with the histogram of the genesis frequency (gray shading, yr^{-1}). The contours of mean GPI (Fig. 5.15a) are able to capture the main features evident in the pattern of genesis occurrence, including centering the enhanced potential for genesis activity within the region of highest SSTs. In particular the extent of the southern and eastern boundaries of the genesis region, along with the asymmetric elongation towards the northeast quadrant are predicted by the GPI. An area where the GPI appears to under-predict the genesis is for the relatively narrow band of activity extending west from the peak genesis region $[\theta_0 < \theta < \theta_0 + 5^\circ, -\pi < \hat{\lambda} < -\pi/4]$, which then manifests as an overall under-prediction in the meridional-mean profile of GPI for longitudes west of the peak genesis, and as a northern shift in the zonal-mean GPI profile.

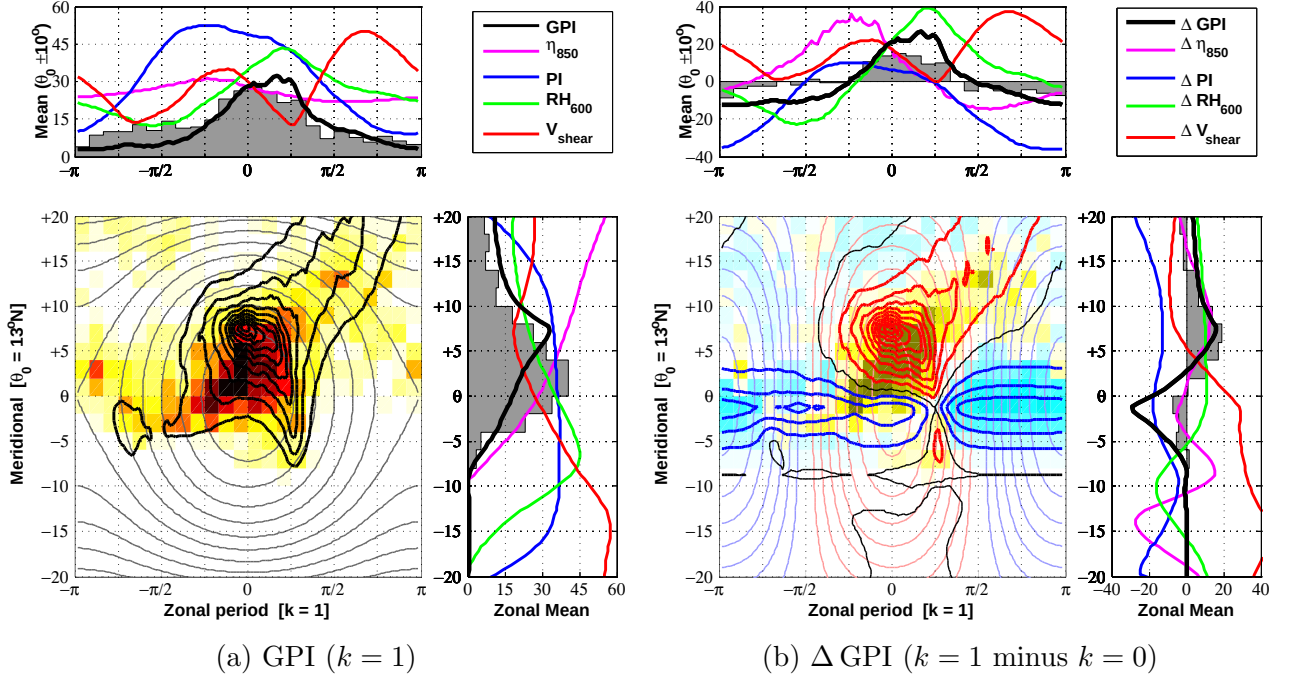


Figure 5.15: Maps showing the genesis frequency (shading) for the $k = 1$ SST perturbation simulation at $\theta_0 = 13^\circ$, overlaid with light contours showing the SST field and thick contours showing the corresponding GPI field. The meridional-mean ($\theta_0 \pm 10^\circ$) and zonal-mean profiles of each of the climatological fields (listed) are shown above and to the right of each panel, respectively. The gray shading shows the histogram of the genesis frequency (yr^{-1}). For the difference field (b), the thin red (blue) contours (± 0.2 K intervals) denote positive (negative) RSST, the thick red (blue) contours denote positive (negative) GPI anomalies, and yellow (blue) shading denote increasing (decreasing) TC occurrence.

The change in the $k = 1$ pattern of GPI from the zonally-symmetric control simulation ($k = 0$) is shown in Figure 5.15b. The red (positive) and blue (negative) contours can be compared to the pattern of increase (yellow shading) and decrease (blue shading) in the actual density of genesis occurrence. The black line in the upper panel shows the difference in the meridional mean GPI ($\theta_0 \pm 10^\circ$), and the right panel shows the corresponding difference in the zonal mean. The gray shading shows the associated change in the genesis frequency. The red contours show a marked increase in GPI between $\sim 2^\circ$ - 12° poleward of the latitude of the maximum SST, and centered in longitude at the peak of the zonal SST perturbation. This region of enhanced GPI coincides well with the yellow shading, showing good agreement with the region

where additional TCs form in the $k = 1$ simulation. The region of enhanced GPI extends poleward and eastward throughout the northeast quadrant of the positive SST perturbation, and is also apparent in the shading of actual occurrence, with enhanced TC genesis extending 10° - 20° poleward and approximately a quarter of a period east of the maximum SST perturbation. The plot also shows that the GPI is over-predicting the reduction in genesis potential in the region immediately south and west of the maximum RSST.

Mean and difference profiles for each of the four climatological fields that impact the GPI (η_{850} , PI, RH_{600} , V_{shear}) are shown by the different colored lines in Figure 5.15. Maps showing more detailed patterns of the individual climatological fields are displayed in the four panels of Figure 5.16 (mean fields) and Figure 5.17 (difference fields). Firstly we note that the absolute vorticity (Fig. 5.16a) increases strongly with latitude due to the contribution of planetary vorticity ($f \sim \sin \theta$). The contribution from changes in the 850-hPa relative vorticity (Fig. 5.17a) is much smaller ($\sim 10\%$), and asymmetrical with respect to the RSST maximum. Here we note that the time-mean relative vorticity is enhanced to the west and north of the RSST maximum, and reduced to the east and south of the RSST maximum.

Secondly, we see that the potential intensity (Fig. 5.16b, 5.17b) generally increases over the region of positive RSST, with its maximum slightly west ($\hat{\lambda} \approx \pi/4$) of the peak SST. While related to the SST, recall that the potential intensity is computed over the entire vertical thermodynamic profile, and so differences between the two-dimensional pattern of RSST and PI reflect differences in the vertical temperature distribution. From the meridional-mean profile of ΔPI one can see that the fractional increase of PI (over the $\text{RSST} > 0$ region) is much smaller than the decrease over the equivalent $\text{RSST} > 0$ region. Hence the zonal-mean profile of ΔPI shows that the total impact from an undulating SST perturbation is to reduce the mean PI across the tropical region.

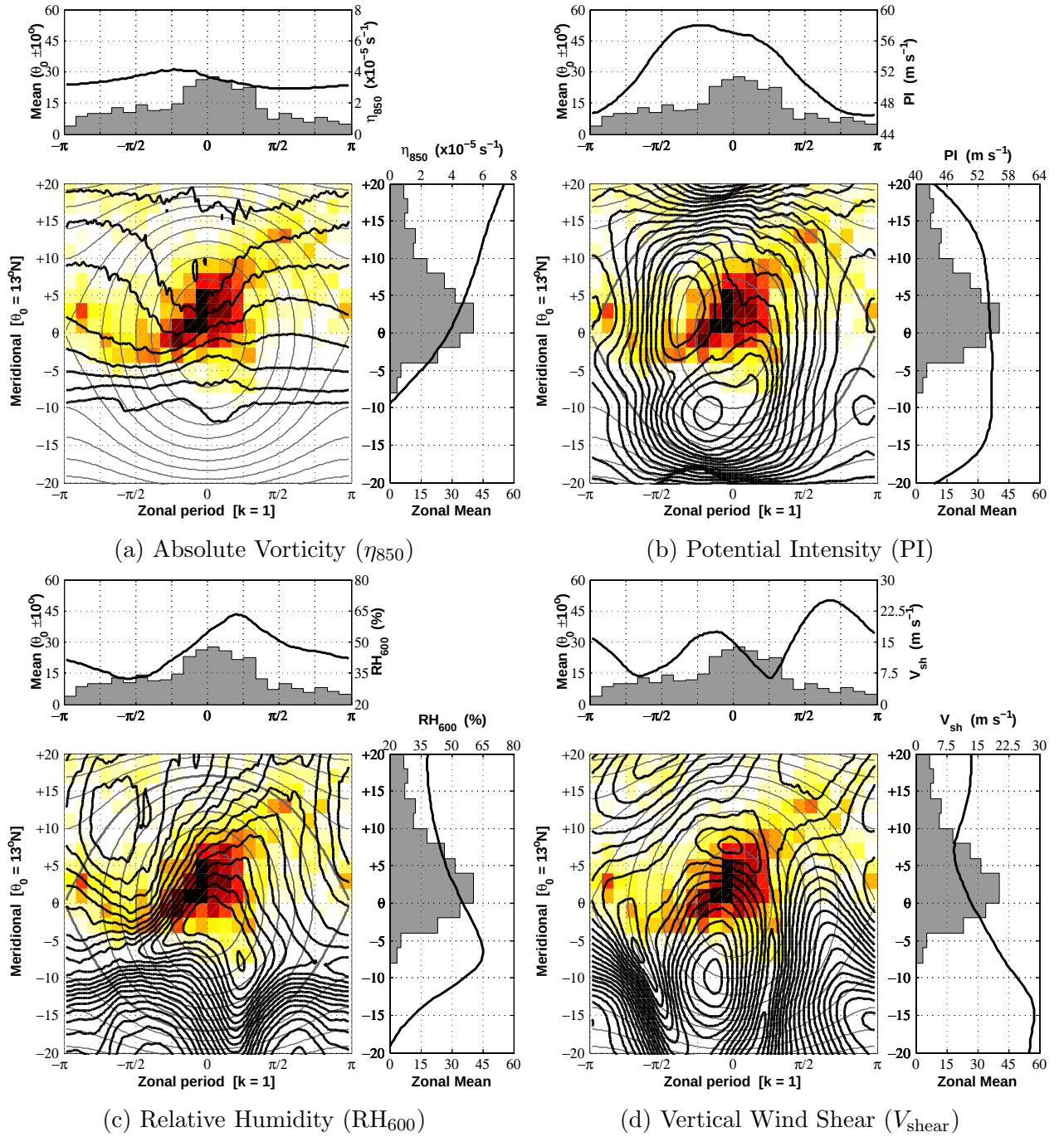


Figure 5.16: Maps showing the genesis frequency (shading) for the $k = 1$, $\theta_0 = 13^\circ$ SST perturbation simulation, overlaid with gray contours showing the SST field (0.5° C contour intervals), and black contours showing the corresponding climatological fields of: a) η_{850} ($1.0 \times 10^{-5} \text{ s}^{-1}$ c.i.), b) PI (1.5 m s^{-1} c.i.), c) RH_{600} (4% c.i.), and d) V_{shear} (2.5 m s^{-1} c.i.). The meridional-mean ($\theta_0 \pm 10^\circ$) and zonal-mean profiles of each field are shown above and to the right of each panel, respectively. The gray shading shows the histogram of the genesis frequency (yr^{-1}).

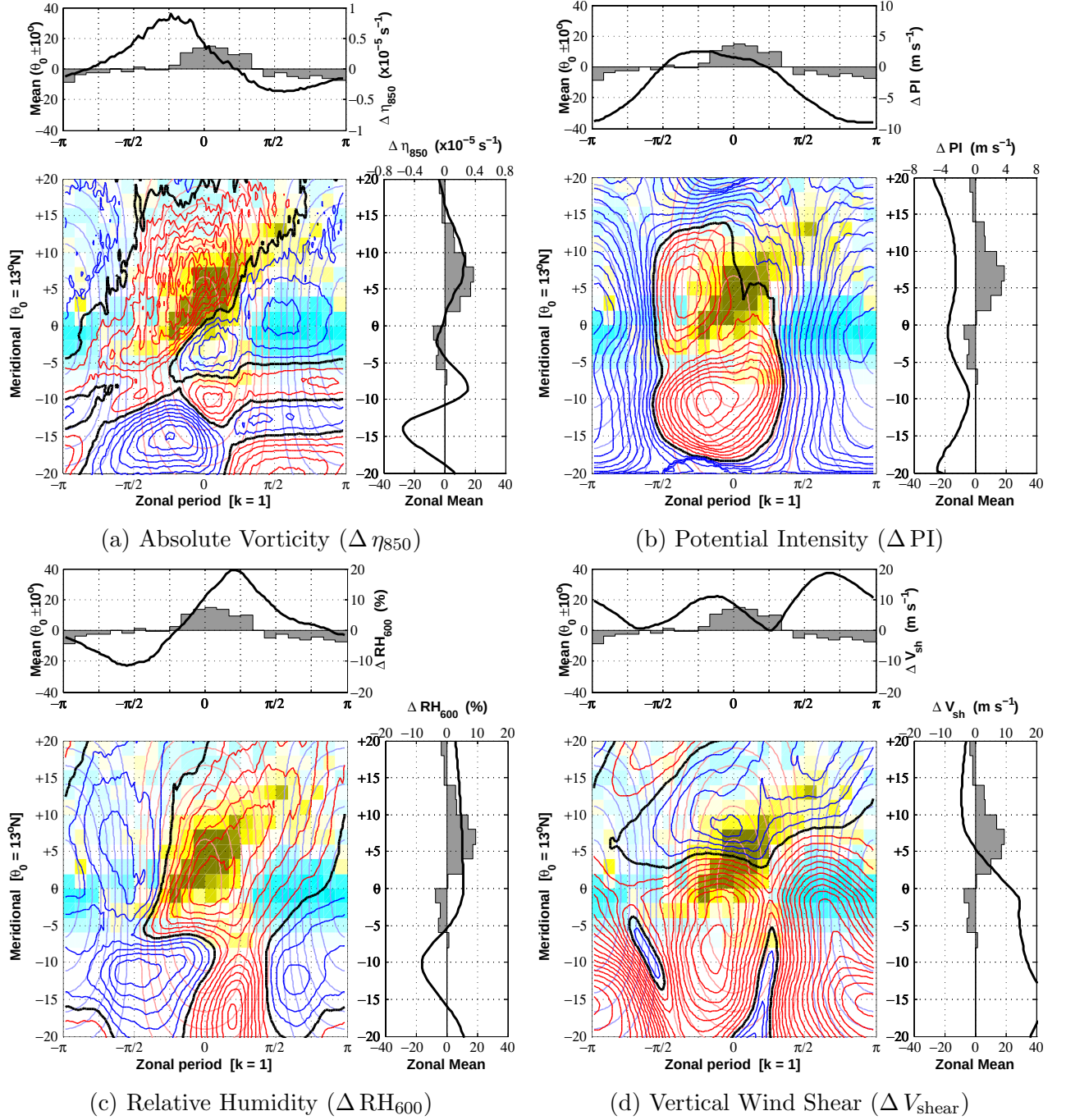


Figure 5.17: Maps showing the increase (yellow shading) and decrease (blue shading) in the genesis frequency for the $k = 1$, $\theta_0 = 13^\circ$ SST perturbation (indicated by the gray contours) simulation when compared to the zonally-symmetric control run. Red (positive) and blue (negative) contours show the corresponding change in climatological fields: a) $\Delta \eta_{850}$ ($\pm 2.5 \times 10^{-6} \text{ s}^{-1}$ contour intervals), b) ΔPI ($\pm 1 \text{ m s}^{-1}$ c.i.), c) ΔRH_{600} ($\pm 4\%$ c.i.), and d) ΔV_{shear} ($\pm 2.5 \text{ m s}^{-1}$ c.i.). Profiles showing the difference in the meridional-mean ($\theta_0 \pm 10^\circ$) and zonal-mean fields are shown above and to the right of each panel, respectively. The gray shading shows the associated change in the genesis frequency (yr^{-1}).

The 600-hPa relative humidity (Fig. 5.16c, 5.17c) is distinctly asymmetrical about the maximum SST, showing a marked reduction in RH to the west of the peak, and a commensurate increase to the east. The minimum in the vertical wind shear pattern shows a semi-annular structure away from the center of the SST maximum, extending from $(\hat{\lambda}, \theta) \approx (-\pi/4, 0) \rightarrow (0, 8) \rightarrow (\pi/4, 0)$ (Fig. 5.16d). The presence of a zonally-asymmetric SST perturbation leads to a dramatic increase the vertical wind shear almost everywhere south of θ_0 , with a slight reduction to the north (Fig. 5.17d).

The structure of these four patterns gives some insight into the pattern of genesis observed in the zonally-asymmetric simulations. The main genesis region is associated with increasing ζ_{850} , increasing PI, increasing RH_{600} . The north-south dipole seen in ΔV_{shear} , along with the asymmetry in ζ_{850} and RH_{600} help to explain the elongation of the favorable genesis region towards the northeast. One might anticipate the reduction in PI and RH_{600} to the west of the peak RSST (especially west of $\hat{\lambda} \approx -\pi/2$), coupled with a slight increase in the vertical wind shear, would suppress the genesis more in this region. However, the slight increase in $\zeta \sim dU/dy$ within this region may play an important role in triggering the disturbances in the first place.

GPI pattern for the $k = 3$ perturbation

The mean GPI for the $k = 3$ SST perturbation simulation (with $\theta_0 = 13^\circ \text{N}$) is shown by the thick black contours in Figure 5.18a, overlaid on the corresponding shaded pattern of TC occurrence (shown earlier in Figure 5.3). The change in GPI between the $k = 1$ (Fig. 5.15a) and $k = 3$ (Fig. 5.18a) SST perturbation simulations is shown in Figure 5.18b. Regions of increasing or decreasing TC occurrence density are shaded yellow or blue, respectively. Also plotted are the mean changes in the four environmental variables of interest ($\Delta\eta_{850}$, ΔPI , ΔRH_{600} , ΔV_{shear}), with more detailed patterns of the individual changes provided in the four panels of Figure 5.19.

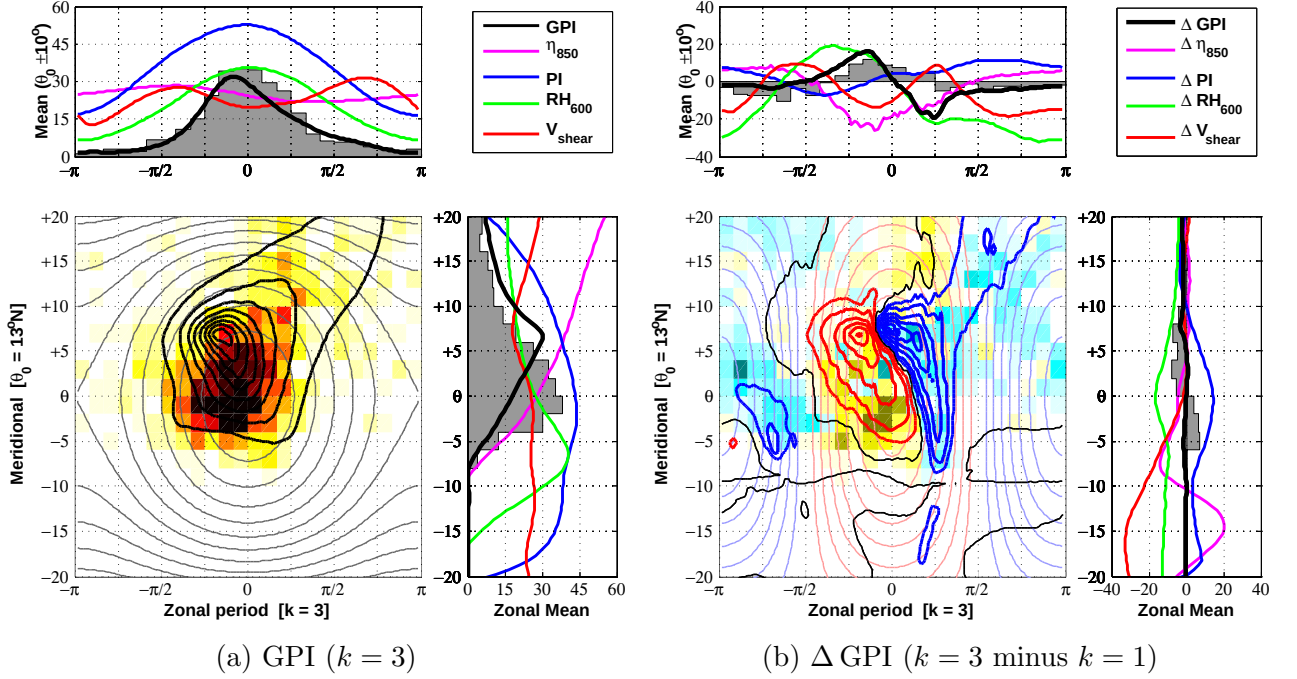


Figure 5.18: Maps showing the genesis frequency (shading) for the $k = 3$ SST perturbation simulation at $\theta_0 = 13^\circ$, overlaid with light contours showing the SST field and thick contours showing the corresponding GPI field. The meridional-mean ($\theta_0 \pm 10^\circ$) and zonal-mean profiles of each of the climatological fields (listed) are shown above and to the right of each panel, respectively. The gray shading shows the histogram of the genesis frequency (yr^{-1}). For the difference field (b), the thin red (blue) contours ($\pm 0.2 \text{ K}$ intervals) denote positive (negative) RSST, the thick red (blue) contours denote positive (negative) GPI anomalies, and yellow (blue) shading denote increasing (decreasing) TC occurrence.

The general pattern of genesis is well captured by the GPI contours (Fig. 5.18a), despite the location of the GPI peak being a little further north and slightly west of the actual peak in genesis, which resides over the maximum RSST. The meridional-mean thermodynamic fields (PI, RH_{600}) are highly correlated with the mean RSST, having their maximum also at $\hat{\lambda} = 0$ and decreasing symmetrically towards the east and west. This is in contrast to the $k = 1$ meridional-mean thermodynamic profiles, where PI had its maximum to the west of the RSST maximum (Fig. 5.16b), and RH increased towards the east (Fig. 5.16c).

The shading in Figure 5.18b shows the contraction of the TC genesis region that has been noted throughout these results. Blue shading over the cool SST anomalies

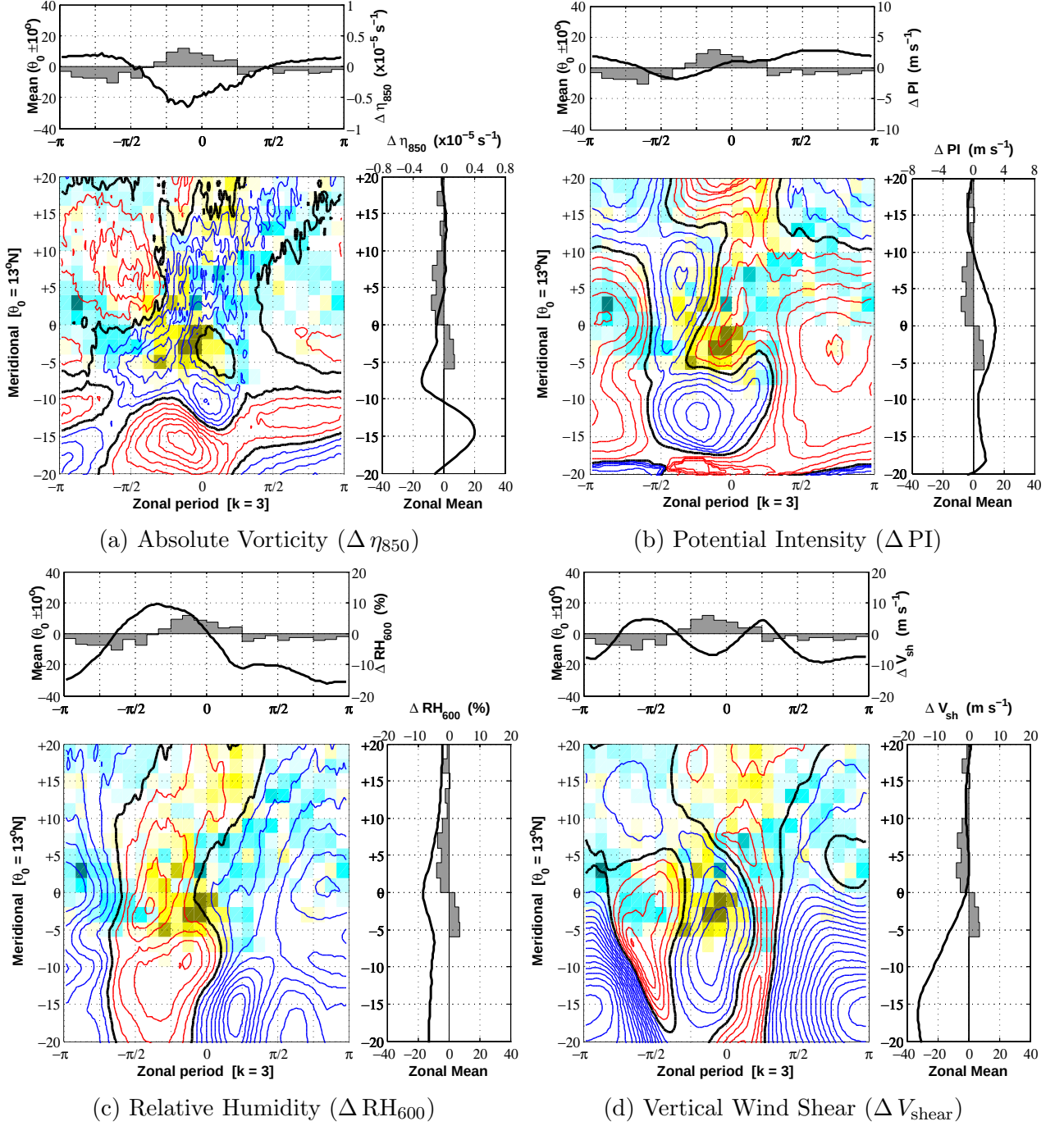


Figure 5.19: Maps showing the increase (yellow shading) and decrease (blue shading) in the genesis frequency for the $k = 3$, $\theta_0 = 13^\circ$ simulation compared to the $k = 1$, $\theta_0 = 13^\circ$ simulation. Red (positive) and blue (negative) contours show the corresponding climatological fields of: a) $\Delta \eta_{850}$ ($\pm 2.5 \times 10^{-6} \text{ s}^{-1}$ contour intervals), b) ΔPI ($\pm 1 \text{ m s}^{-1}$ c.i.), c) ΔRH_{600} ($\pm 4\%$ c.i.), and d) ΔV_{shear} ($\pm 2.5 \text{ m s}^{-1}$ c.i.). Profiles showing the difference in the meridional-mean ($\theta_0 \pm 10^\circ$) and zonal-mean fields are shown above and to the right of each panel, respectively. The gray shading shows the associated change in the genesis frequency (yr^{-1}).

show that the occurrence of TCs over these cooler regions is scarcer in the $k = 3$ simulation. Yellow shading within the core region of the warm RSST anomaly confirms the increase in genesis occurrence density with increasing k . The change in occurrence is predicted by the increase and decrease in GPI shown by the red and blue contours; the thin black contour is plotted to show the $\Delta \text{GPI} = 0$ contour. The area of $\Delta \text{GPI} > 0$ is mostly confined to relatively small region over the warmest values of RSST, showing greater genesis potential within the higher RSST contours. This is contrasted with the broad regions of suppressed genesis potential that encroach into the flanks of the $\text{RSST} > 0$ region, acting to narrow the region of GPI in the $k = 3$ simulation compared with the $k = 1$ experiment.

We next look at Figure 5.19 to analyze changes in the climatological fields important for mediating genesis potential, in order to relate these changes to the observed contraction and intensification of the region of TC genesis activity. Firstly we note that ζ_{850} decreases throughout the genesis region, except further northwest of the RSST maximum (Fig. 5.19a). Since this reduces genesis potential across the region we are unable to attribute the observed changes in the TC genesis pattern to changes in vorticity. We next consider the potential intensity (Fig. 5.19b), and here we see a semi-annular structure of reduced PI extending from the northwest flank around through the southwest and southern flank of positive RSST values, acting to reduce genesis potential over these outer regions. Furthermore, a small region showing a marked increase in PI is located over the RSST maximum, precisely in the region of increased TC genesis. The changes in RH_{600} show an increase in relative humidity to the west of the RSST maximum and similar decrease to the east (Fig. 5.19c). These changes bring greater symmetry (about the RSST maximum) to the pattern of relative humidity (recall Fig. 5.16c, 5.17c).

Finally we look at the pattern of change in the time-mean vertical wind shear (Fig. 5.19d). The pattern shows a significant increase in vertical wind shear

($\sim 5\text{--}10\text{ m s}^{-1}$) over the flanks of the positive RSST regions, precisely where the genesis occurrence is seen to reduce. Moreover, a decrease in shear is seen over the narrow central region of high RSST. It is likely that this changing pattern of vertical wind shear is a major reason why the genesis region contracts with increasing perturbation wavenumber k , supported by changes in the thermodynamic fields that bring their maxima in alignment with the maximum RSST.

Magnitude of GPI over the genesis region

Applying the same approach developed in the previous section (Sec. 5.2.4), the mean GPI is next computed over two different regions. Firstly, the $\langle \text{GPI} \rangle_{\text{RSST} > 1}$ for each zonally-asymmetric simulation is computed over the fixed tropical region(s) with $\text{RSST} > +1\text{ K}$, corresponding to the warmest $\sim 10\%$ of the tropical area ($\theta_0 \pm 30^\circ$). Secondly, the $\langle \text{GPI} \rangle_{50\%}$ is computed over the area of the RSST contour containing 50% of TC genesis locations within each simulation. The density of TC occurrence for each simulation is plotted against these two measures of $\langle \text{GPI} \rangle$ in Figures 5.20 and 5.21.

The magnitude of $\langle \text{GPI} \rangle_{\text{RSST} > 1}$ increases by $\sim 10\%$ as the zonally-asymmetric patterns are shifted poleward by 3° (Fig. 5.20, blue $\rightarrow$ pink markers), with the ordinate showing a corresponding increase in TC occurrence density. This increase in TC occurrence density (with θ_0) is consistent with the general pattern of increasing genesis frequency accompanying a poleward shift in the latitude of the maximum SST (Sec. 3.2, 5.2.1). While $\langle \text{GPI} \rangle_{\text{RSST} > 1}$ is thus able to predict the increase in frequency over a fixed region with increasing θ_0 , it is unable to predict the significant increase in occurrence density that is associated with increasing k .

From Figure 5.21 we see a strongly positive and compact quasi-linear relationship between the predicted genesis potential $\langle 5 \rangle_{\text{G}} 0\text{GPI}$, and the resulting occurrence density. As the zonal wavenumber of RSST increases from $k = 1$ to $k = 3$ the magnitude

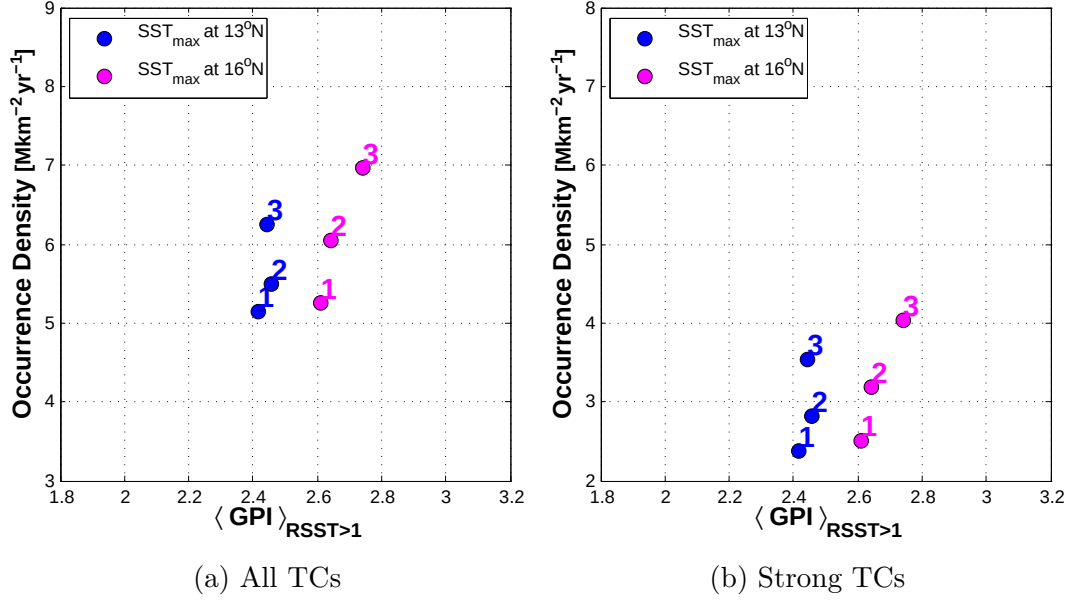


Figure 5.20: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean Genesis Potential Index ($\langle \text{GPI} \rangle_{\text{RSST} > 1}$), computed over the region(s) where $\text{RSST} > +1$ K. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

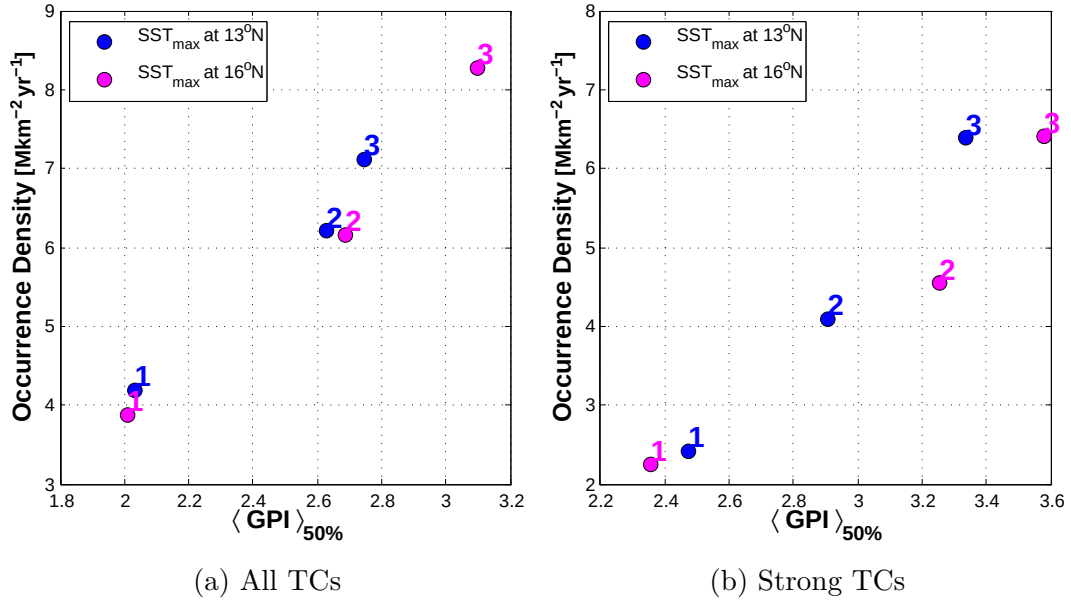


Figure 5.21: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean Genesis Potential Index ($\langle \text{GPI} \rangle_{50\%}$), which has been computed over the region(s) enclosed by the RSST contour containing 50% of TC genesis locations within that simulation. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

of $\langle \text{GPI} \rangle_{50\%}$ increases by $\sim 50\%$ and the occurrence density doubles. By computing the mean GPI over the regions in the simulation having the same proportion of resulting TCs, this measure better samples the environmental conditions coincident with TC genesis.

The changes in the individual components of the GPI can be analyzed in a similar way, and Figure 5.22 shows the $\langle \eta_{850} \rangle_{50\%}$, $\langle \text{PI} \rangle_{50\%}$, $\langle \text{RH}_{600} \rangle_{50\%}$, and $\langle V_{\text{shear}} \rangle_{50\%}$ plotted against the TC occurrence density. The mean potential intensity (Fig. 5.22b) and relative humidity (Fig. 5.22c) over the genesis region increase with increasing k , especially moving from $k = 1$ to $k = 2$. The mean vertical wind shear (Fig. 5.22d) decreases with increasing k , most notably between $k = 2$ and $k = 3$. Hence these three fields contribute to the strong relationship observed between $\langle \text{GPI} \rangle_{50\%}$ and TC occurrence. The mean absolute vorticity (Fig. 5.22a) shows little change with k , except in the $\theta_0 = 13^\circ \text{ N}$ simulations where a small decrease is seen with increasing k .

5.3 Summary

In this chapter the set of fixed-SST aquaplanet experiments have been expanded to include prescribed perturbations to the zonal pattern of SST. The zonal asymmetry introduced was a simple cosine wave, having wavenumber 1, 2 or 3. The wavenumber set the number of zonal SST maxima and thus also varied the mean magnitude of the zonal SST gradient. Exploring idealized warm and cool zonal asymmetries is a natural step towards understanding the impact of large-scale basin-to-basin SST variability on global TC activity and was motivated in part by the apparent utility of relative SST (the local SST relative to the zonal mean) in moderating tropical cyclone activity.

As expected, compared to the zonally-symmetric control experiments, the tropical cyclones that were simulated by the model with a zonally-asymmetric SST had

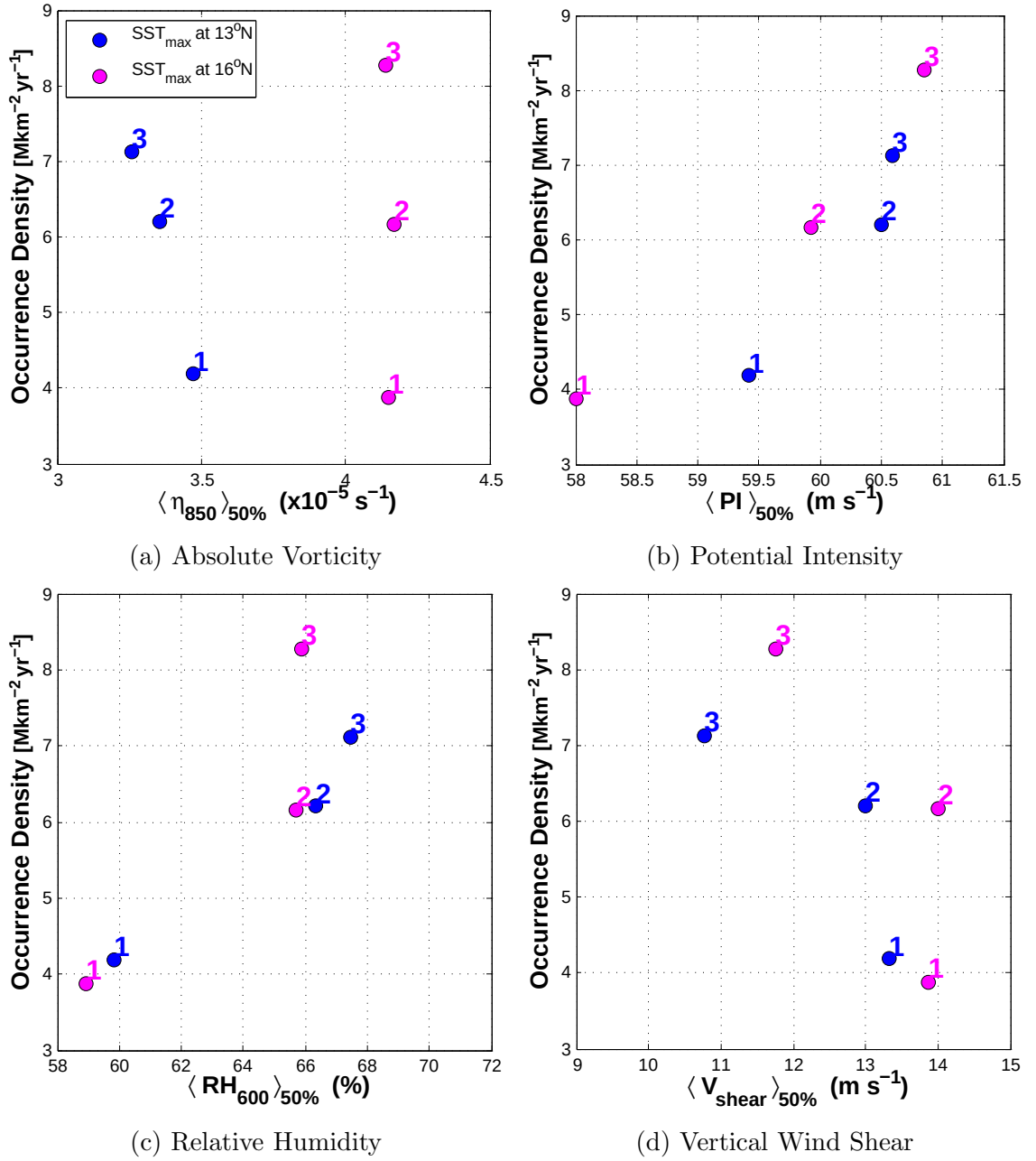


Figure 5.22: TC genesis occurrence density (TCs per 10^6 km^2 per year) plotted against the time-mean fields of η_{850} , PI, RH_{600} , and V_{shear} , which have been spatially averaged over the region(s) enclosed by the constant RSST contour containing 50% of TC genesis locations. Each marker corresponds to a different zonally-asymmetric simulation (with colors indicating θ_0 , and numbers indicating k).

their genesis occur predominantly over the regions of positive SST perturbation, with fewer TCs forming in the tropical regions with cool SST anomalies. As the zonally-asymmetric SST perturbation wavenumber increased from 1 to 3 there was little change in the global frequency of TC genesis, however the geographical regions where genesis occur contracted, and the density of TC genesis within those regions increased.

The change in total genesis frequency moving from the $k = 0$ to $k \geq 1$ simulations depends on the latitude of the maximum SST of the control climate. When the maximum SST is close to the equator the introduction of a zonally-asymmetric perturbation increases the global genesis, due to the accompanying northward shift of the localized genesis region and increase in absolute vorticity. However, when the latitude of the maximum SST is shifted further poleward, the zonally-asymmetric perturbation leads to a reduction in the total number of TCs, most likely due to the dominating influence of enhanced vertical wind shear.

The geographical region(s) of favored TC genesis can be approximated by contours of perturbation sea surface temperature that contain a certain percentage of the TC genesis locations within an individual simulation. When the zonally-asymmetric SST perturbation wavenumber is increased from 1 to 3, corresponding to a tripling of the average magnitude of the zonal gradient of SST, the geographical area of TC genesis contracts, with genesis occurring (on average) over higher values of perturbation SST. The reduction in the occurrence of TC genesis on the flanks of the warm region(s) appears to be related to the increased vertical wind shear associated with the different SST patterns. An increase in the average magnitude of the zonal gradient of SST results in an increase in the simulated TC genesis frequency per unit area within the favored genesis region(s). The increase in TC frequency is consistent with the colocated increase in mean precipitation rate and upward mass flux over the region.

Chapter 6

Conclusion

6.1 TCs in a zonally-symmetric aquaplanet GCM

In the zonally symmetric, interhemispherically asymmetric climates of the HiRAM aquaplanet presented in this study, tropical cyclones form on the poleward flank of the off-equatorial ITCZ, subsequently strengthening for a while before reaching a maximum intensity and then decaying. The evolution of the TC environment during this post-genesis poleward migration to higher latitudes is characterized by decreasing SSTs, decreasing Potential Intensity (closely tied to the decreasing SSTs), decreasing vertical motion (moving away from the ITCZ region into the subsidence of the extra-tropics), and increasing vertical wind shear.

When analyzing the factors impacting TC activity in these simulations it is inherently difficult to separate any influences from the prescribed SST profile and the latitude of the ITCZ; the latter is of course related to the former in some nontrivial way. However, by fixing some aspect of the SST profiles between experiments (e.g., the magnitude of the maximum SST and ∇ SST) while adjusting some other characteristic (e.g., the latitude of the maximum SST), a set of aquaplanet climatologies were formed, each with a small perturbation to the SST profile and associated changes

in the latitude of the ITCZ. The resulting variation in the genesis latitude, frequency and intensity have been investigated across a broad range of zonally-symmetric environments in order to identify consistent and robust changes in the statistics of the simulated TCs.

6.1.1 Summary of major results

The major findings from this study of simulated tropical cyclone activity in the Hi-RAM aquaplanet with zonally-symmetric SSTs are briefly summarized as follows:

1. **Tropical cyclones prefer to form near 15° N latitude:** With a sufficiently flat meridional SST profile, tropical cyclones tend to prefer to form at $\sim 14.5^\circ$ N in these simulations. As the meridional SST gradient is increased, both the ITCZ and genesis region shift towards the latitude of the maximum SST.
2. **Tropical cyclone activity increases as the latitude of the maximum sea surface temperature shifts poleward:** The total accumulated cyclone energy from the simulated TCs increases as the latitude of the maximum SST is shifted poleward, due in part to an increase in the actual frequency of TC genesis, and also due to an increase in the average intensity of the TCs.
3. **Tropical cyclone genesis frequency increases as the latitude of the ITCZ shifts poleward:** The experiments which move the ITCZ further poleward while holding the latitude of the maximum SST fixed (through changes in the magnitude of ∇SST) result in a higher genesis frequency. These experiments also reveal that the TCs which form in environments where the ITCZ is further poleward and closer to the latitude of the maximum SST are on average less intense, and hence the total activity is relatively insensitive to the changes in the latitude of the ITCZ alone.

4. **Tropical cyclone genesis frequency is related to both the upward mass flux and ambient vorticity of the large-scale environment:** While the strength of mid-level ascent ($|\omega_{500}|$) has been shown to be an important mediator of TC frequency, the upward mass flux over the genesis region alone is not determinative of TC frequency, as notably demonstrated in the slab ocean experiments. Changes in the latitude of the genesis region, and thus the absolute vorticity available to developing TCs, also influence the genesis frequency. Most interesting, the strength of mid-level ascent at $\sim 15^\circ$ N appears to be tightly related to the TC genesis frequency across all experiments.

5. **Tropical cyclone intensity decreases as the meridional gradient of sea surface temperature increases:** The average accumulated cyclone energy per TC decreases as the meridional SST gradient increases, due in part to a reduction in the average duration, and also due to a reduction in the average maximum wind speed of the TCs within the simulations. Recognizing that TC intensities in this model are a function of resolution, one must assume that results on statistics such as the accumulated cyclone energy will change quantitatively at higher resolution. Whether the qualitative results described here will change at higher resolution remains to be seen.

6. **Tropical cyclones acquire a higher lifetime maximum intensity at latitudes that are further poleward:** We find a positive relationship between the mean lifetime maximum intensity of TCs and the corresponding mean latitude at which the maximum occurs. There is also a strong correlation between the mean lifetime maximum intensity of TCs and the period of time that the TCs undergo intensification. These two aspects of the TC tracks (the latitude of maximum intensity, and the period until maximum intensity) are closely related, and it would be interesting to investigate this inter-dependence further.

6.2 TCs in a zonally-asymmetric aquaplanet GCM

The set of fixed-SST aquaplanet experiments were expanded to include prescribed perturbations to the zonal pattern of SST. The zonal asymmetry introduced was a simple cosine wave, having wavenumber 1, 2 or 3. The wavenumber set the number of zonal SST maxima and thus also varied the mean magnitude of the zonal SST gradient. Exploring idealized warm and cool zonal asymmetries is a natural step towards understanding the impact of large-scale basin-to-basin SST variability on global TC activity and was motivated in part by the apparent utility of relative SST (the local SST relative to the zonal mean) in moderating tropical cyclone activity.

6.2.1 Summary of major results

The major findings from this study of simulated tropical cyclone activity in the Hi-RAM aquaplanet with zonally-asymmetric SSTs are briefly summarized as follows:

1. **Tropical cyclones preferentially form over tropical regions where the sea surface temperature is greater than the zonal mean:** As expected, compared to the zonally-symmetric control experiments, the tropical cyclones that were simulated by the model with a zonally-asymmetric SST had their genesis occur predominantly over the regions of positive SST perturbation. Fewer TCs formed in the tropical regions with cool SST anomalies.
2. **The total (global) frequency of tropical cyclone genesis may either increase or decrease with the introduction of a zonally-asymmetric perturbation pattern in sea surface temperature:** The change in total genesis frequency depends on the latitude of the maximum SST of the control climate. When the maximum SST is close to the equator the introduction of a zonally-asymmetric perturbation increases the global genesis, due to the associated poleward shift of the genesis region and increase in absolute vorticity.

However, when the latitude of the maximum SST is shifted further poleward, the zonally-asymmetric perturbation leads to a reduction in the total number of TCs, likely due to the dominant influence of enhanced vertical wind shear.

3. **The total (global) frequency of tropical cyclone genesis is relatively insensitive to the wavenumber of the zonal sea surface temperature perturbation:** An increase in the zonally-asymmetric SST perturbation wavenumber from 1 to 3 resulted in relatively small, nonmonotonic changes in the total number of tropical cyclones in each simulation.
4. **The geographical region of tropical cyclone genesis contracts as the mean magnitude of the zonal gradient of sea surface temperature increases:** The geographical region(s) of favored TC genesis can be approximated by contours of perturbation sea surface temperature that contain a certain percentage of the TC genesis locations within an individual simulation. When the zonally-asymmetric SST perturbation wavenumber is increased from 1 to 3, corresponding to a tripling of the average magnitude of the zonal gradient of SST, the geographical area of TC genesis contracts, with genesis occurring (on average) over higher values of perturbation SST. The reduction in the occurrence of TC genesis on the flanks of the warm region(s) appears to be related to the increased vertical wind shear associated with the different SST patterns.
5. **The occurrence density of tropical cyclones within the genesis region increases as the mean magnitude of the zonal gradient of sea surface temperature increases:** An increase in the average magnitude of the zonal gradient of SST results in an increase in the simulated TC genesis frequency per unit area within the favored genesis region(s). The increase in TC frequency is consistent with the colocated increase in mean precipitation rate and upward mass flux over the region.

6.3 In closing

The aquaplanet configuration of HiRAM or other atmospheric general circulation models with comparable resolution provides an ideal testbed in which to continue to explore the sensitivity of genesis to various aspects of the model’s climatology. Although beyond the scope of this investigation, convectively-coupled waves that propagate throughout the tropical region, along with instabilities that develop along the ITCZ itself are likely important in the formation of the precursor disturbances, from which a certain fraction develop into self-sustaining tropical cyclones. The extent to which the large-scale influences the fraction of disturbances that develop into tropical cyclones still remains an open question.

In an early review of the theory of tropical cyclone formation, Yanai (1964) began by asking “What produces a hurricane or a typhoon? Despite years of effort by many meteorologists, the question has not yet been completely answered...”. Fifty years later and there have been staggering advances in the platforms available for routinely detecting and observing tropical storm formation, accompanied by great progress in understanding the kinetic and thermodynamic aspects of development, and dramatic improvement in the computational capability of simulating these processes in complex numerical models. However, we must similarly conclude with Yanai that despite years of effort there remain many very important, incredibly interesting, and yet unsolved questions surrounding the environmental controls on tropical cyclone formation, especially when it comes to projecting future changes in tropical cyclone activity in different climates. The question of what produces a tropical cyclone has not yet been completely answered.

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