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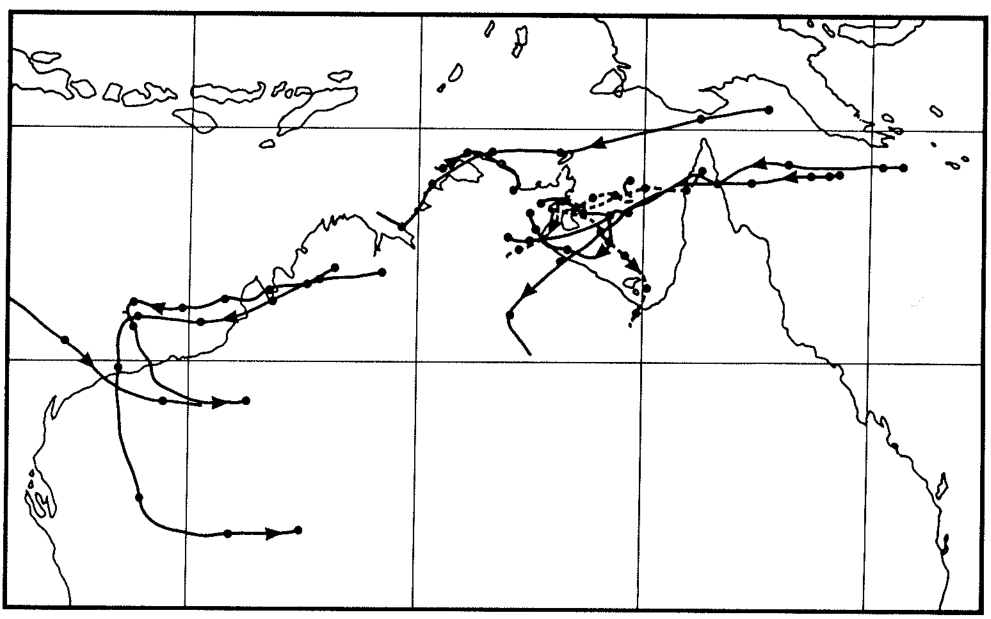
J.L. Evans



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Envisaged Impacts of Enhanced Greenhouse Warming on Tropical Cyclones in the Australian Region

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Abstract

Changes in the characteristics of tropical cyclones due to enhanced greenhouse warming have been foreshadowed in both the popular and scientific literature. This report presents an overview of the current theories relating to tropical cyclones and uses this information to deduce potential changes in tropical cyclone characteristics due to enhanced greenhouse warming. The main conclusion is that the more sensational claims of increased tropical cyclone strength and occurence have not considered the majority of the environmental factors that affect tropical cyclones. Further research is neccessary to explore the impacts of these environmental factors on tropical cyclone behaviour.

1. Introduction

Impacts of the changes envisaged in large–scale atmospheric circulation due to increasing greenhouse gases have been widely discussed in recent times, both in the scientific literature and in the media. Changes in the characteristics and distribution of tropical cyclones have been considered to be among the possible effects of the proposed changes in the structure of the atmosphere. Indeed, forecasts of global increases in the number and intensity of tropical cyclones have been put forward (Emanuel, 1987) and widely reported in the popular literature (The Age, 1990; The Sunday Age, 1990; Greenpeace, 1989; The Australasian Post, 1989; The Dominion, 1988; The Advertiser, 1988; The News, 1988; The Australian, 1987). The spectre of tropical cyclones intruding into the heart of the Australian deserts and as far south as Sydney and Perth has also been raised. However, the evidence for these claims has not been well documented.

The aim of this report is to evaluate our current knowledge of tropical cyclones and then to use this knowledge to examine the possible effects on tropical cyclones, especially in the Australian region, of changes to the general circulation in a world of doubled CO_2, as foreseen by six well–documented General Circulation Model (GCM) simulations. Further, an attempt is made to identify areas where our knowledge of tropical cyclones still is very limited.

The approach taken here is to infer modifications to tropical cyclone behaviour in an environmental of enhanced greenhouse warming. This technique is necessary since GCMs cannot resolve tropical cyclones, so one is unable to investigate the changes in tropical cyclone populations due to the warming envisaged by the model by interrogating the model directly. Two previous studies have addressed this problem. Holland et al (1987) discussed the potential changes to Australian region tropical cyclones based upon the Greenhouse '87 scenario and Raper (1990) has drawn on a world–wide tropical cyclone dataset to investigate observational relationships between climatic changes and severe tropical cyclones. This document draws on both of these papers, but also includes recent theoretical work and an updated greenhouse scenario.

Section 2 describes the 'Greenhouse Scenario' used in this paper and sections 3 through 7 consider current tropical cyclone climatology, El Nino/Southern Oscillation (ENSO) and Quasi Biennial Oscillation (QBO) effects, genesis and intensification, structure and motion, respectively, both in the current climate and that given by the greenhouse scenario adopted. Finally, a summation of the potential impacts of greenhouse gas increase on tropical cyclones is drawn and some areas requiring further study are highlighted.

2. The Greenhouse Scenario Used

The effect of many of the trace gases in the Earth's atmosphere, such as water vapour, carbon dioxide, methane and ozone, is to warm the planet. This warming is known as the 'Greenhouse Effect' and these gases are referred to as greenhouse gases. Some of these gases (for example, carbon dioxide and methane) have been observed to increase substantially in concentration since the industrial revolution (Pearman, 1988; IPCC, 1990). It is the effect of this increase in CO_2 and other gases, the 'Enhanced Greenhouse Effect', that is of concern at the moment. By extrapolation, one would expect an increase in global mean temperature with increased greenhouse gas concentrations. However, the distribution of this temperature change and its magnitude are uncertain. The feedback of this temperature change into the dynamics and thermodynamics of the atmosphere–ocean system is of obvious concern. At present, GCMs are our best tool to investigate the broadscale patterns of this enhanced greenhouse effect (Mitchell, 1989).

In order to discuss the potential implications of changes in tropical cyclone characteristics in a climate modified by increasing concentrations of greenhouse gases, the climate changes envisaged from this enhanced greenhouse effect must first be identified. Temperature and moisture scenarios from five General Circulation Models (GCMs) have been considered (Hansen et al, 1988; Wilson and Mitchell, 1987; Schlesinger and Zhao, 1987; Washington and Meehl, 1984; Manabe and Weatherald, 1980). In the Australian region, Whetton (1990) has shown that the CSIRO4 GCM (Gordon, 1981; Hunt and Gordon, 1989) gives the most reliable representation of the current climate. A greenhouse temperature and moisture scenario based upon the common elements of climate forecast by all of these models has been derived and is summarised below. Ryan et al (1990) have compared the position of the monsoon shear line in the Australian region with the model predictions and find that the CSIRO4 control run agrees reasonably well with their ten year climatology of monsoon mean position and variability. They show that the CSIRO4 model predicts very little change in the mean position of the monsoon trough, although the wind strength may increase somewhat. This study currently provides the best estimate of the behaviour of the monsoon trough in an environment of enhanced greenhouse warming and so it is used here. The position of the monsoon trough is most important for determining tropical cyclogenesis locations (McBride

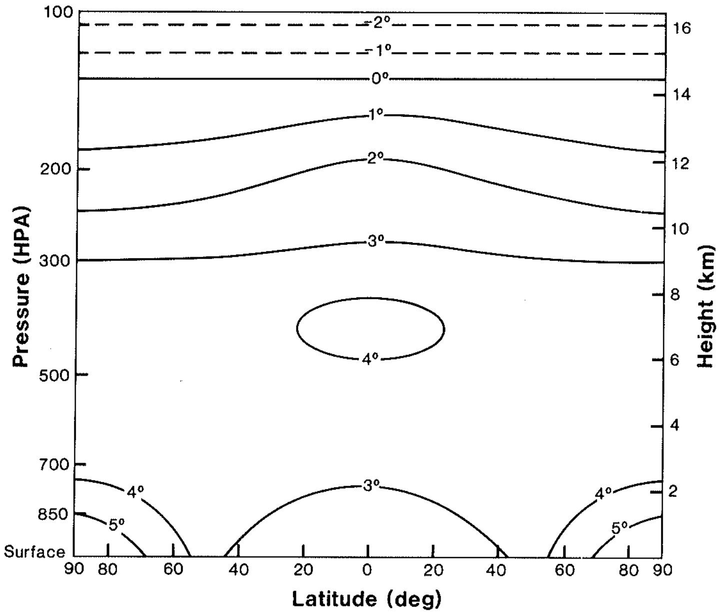

Figure 1: Composite pattern of enhanced greenhouse minus current climate temperature anomalies. This is the basic scenario referred to in this paper.

and Keenan, 1982), so any movement of the monsoon trough should result in a coincident shift in the genesis locations of Australian region tropical storms. This also would impact indirectly on preferred cyclone paths.

It should be noted that this scenario is taken from the best estimates made by the models to date and is not to be taken as a 'forecast' of 'The Greenhouse Climate'. The scenario used in this report is: surface temperature gradients from the equator to poles will be decreased, with the poles warming more than the equatorial regions in the lower levels. This will decrease the horizontal baroclinicity from equator to poles. Coincident with this atmospheric surface temperature warming will be an increase in the sea surface temperatures (SSTs). At the equator, warming will increase up to the middle troposphere, with a maximum increase of about 4° at 400mb. Middle–tropospheric moisture also will increase in the tropics. The combined effect of these changes could be to stabilise the tropics and potentially reduce convective activity. The increased tropospheric temperatures are balanced by a fairly uniform stratospheric cooling from equator to pole. Figure 1 is a schematic of the envisaged temperature changes from the current climate. The average position of the monsoon shear line is not expected to change significantly, although a slight strengthening of the winds in this region is indicated (Ryan et al, 1990).

The models have many approximations and inaccuracies, such as limited horizontal and vertical resolution and simplified treatments of their oceans and convection, for which improvements currently are being devised. Some of these problems are discussed below.

As yet, computational facilities are inadequate for providing fine resolution (on the scale of 100 km or better) general circulation model runs. The models currently in use typically have a horizontal resolution of 5° latitude and 7° longitude

(Schlesinger and Mitchell, 1987; Mitchell, 1989) and so are useful only for temporally and zonally averaged scenarios (i.e. zonal averaged mean January scenarios, for example). Regional-scale variations and realistic synoptic variability must be deduced until GCM resolution can be improved. Typically, the models have between two and nine vertical layers in the atmosphere and resolution of the oceans is generally worse. Until recently only the Goddard Institute of Space Sciences (GISS) and British Meteorological Office (UKMO) models allowed for transport of heat in the oceans (Mitchell, 1989) by what is termed a 'Q-flux' correction. This heat transport is assumed to be the same for both the control ($1 \times CO_2$) and doubled CO_2 cases. The CSIRO4 model has such a coupled 'slab' ocean. These coupled 'oceans' are simple fixed-depth, uniform mixed-layers in all but the GISS model, which prescribes a seasonal variation to the mixed-layer depth. Hence no modification of ocean circulation and potential feedback to the atmospheric system is possible at this stage. Meehl (1990) and Washington (1990) describe recent developments in the NCAR Community Climate Model (CCM), coupling the atmospheric GCM to a realistic ocean model. By considering ocean currents and salinity variations, the CCM allows for more realistic feedbacks due to variable heat transport and wind forcing changes. Major differences with previous work show up in the North Atlantic, with cooling predicted there in the longer term, in spite of an overall global warming (see also Manabe and Stouffer, 1988). Cooling also is observed at high latitudes in the Southern Ocean. This will affect the high latitude warming and meridional temperature gradient scenarios deduced from coupled GCM-slab ocean calculations. A detailed discussion of the impact of more realistic ocean modelling is provided in Section 4 in relation to the El Nino/Southern Oscillation (ENSO) phenomenon.

Parameterisation of clouds and precipitation in the global models is extremely crude at present, providing one of the largest sources of uncertainty in the modelling results. Since the models are unable to resolve such features as tropical storms and fronts, their prediction of clouds only relies on convergence and moisture transport and convective adjustment on the large-scales. Cloud feedbacks to the climate system may provide significant modifications, at least locally, to the current scenarios.

With all of these caveats on the models, they are still our most useful tools for developing scenarios of broadscale pattern variation in a future climate and so are used here. Moreover, at least in a zonally averaged sense, the GCMs produce fairly reasonable 'control' climates (i.e. the currently observed climate), including a realistic annual cycle (Whetton, 1990). Their ability to reproduce this response to large variations in radiative forcing provides some basis for pursuing this line of investigation for the response to greenhouse forcing changes.

3. Tropical Cyclone Climatology

Annual global tropical cyclone activity is remarkably stable. Approximately eighty tropical storms are observed around the globe each year and of these, two thirds reach hurricane intensity (Figure 2). The average annual variation of occurrence is only ±7%, although the regional variations are much wider and are largely uncorrelated from one region to another (Raper, 1990). For example, within the Australian/Southwest Pacific region, the average number of tropical storms observed

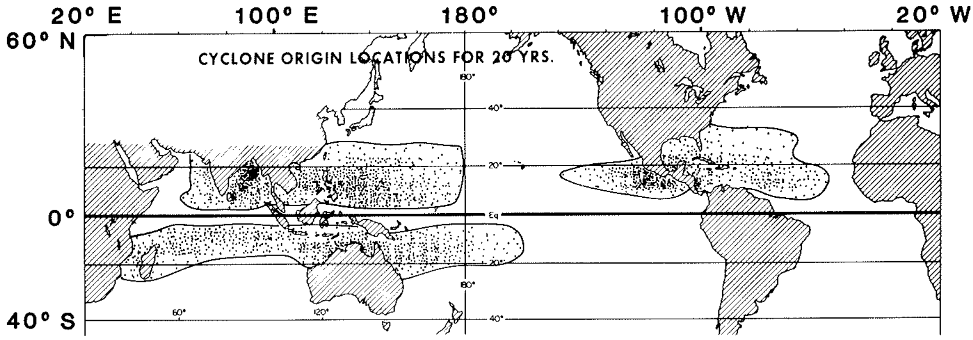

Figure 2: Global tropical cyclogenesis locations for a twenty year period (Gray, 1975).

between 1950 and 1986 was 14.8 with an annual variation of 40 %. Some contributions to this large, regional, interannual variability are discussed in the next section.

As in all other regions, tropical storms in the Australian region show a definite seasonal variation (Bureau of Meteorology, 1978; Lourensz, 1981; McBride and Keenan, 1982; Holland, 1984a and others). The season extends from November to May with a maximum in cyclone activity in January and February. This seasonal bias coincides with the maximum in regional sea surface temperatures (exceeding the critical lower limit of 26.5° C) and a favourable large-scale flow evident by the presence of the monsoon trough (McBride and Keenan, 1982). The implications of this seasonal variation will be discussed further in relation to cyclogenesis in Section 5.

Within the Australian region, tropical cyclones have three preferred areas of formation (Figure 3). Although tropical cyclones in the northern region form over very warm waters, they tend to have shorter lifetimes than in other parts of the Australian region, due to the proximity of land (Holland, 1984a; Holland et al, 1988).

Changes due to greenhouse warming, discussed in Section 2, include an increase in the tropical SSTs and a possible extension of the season in which SSTs equal or exceed the current cyclogenesis threshold of 26.5° C in genesis regions. Coupled with this is the tentative forecast that the tropical atmosphere will have more moisture in the middle levels and the observed thermodynamic conditions for tropical cyclogenesis are satisfied over a wider region. However, the reasons for this lower limit on the sea surface temperature currently are not well understood. Graham and Barnett (1987) have noted that SSTs of approximately 27° C are necessary

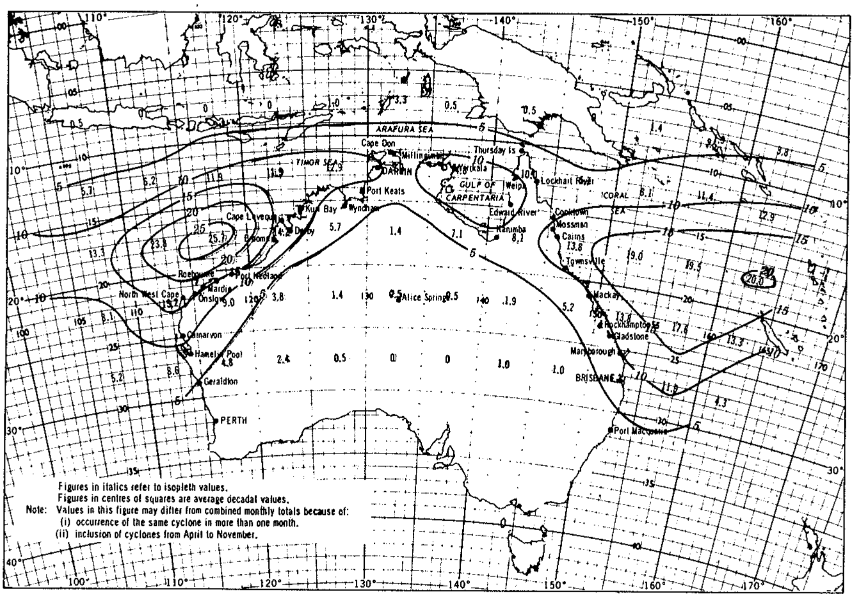

Figure 3: Mean decadal occurrence of tropical cyclones in the Australian region (Lourensz, 1981).

(but not sufficient) for large–scale, deep tropical convection to occur. These warm SSTs must be accompanied by a convergent low–level wind field. Once this SST threshold has been exceeded, no further evidence of a link between SSTs and convection is observable; that is, convection does not appear to increase in intensity with increasingly warmer SSTs. The implications of these links will be discussed further in Section 5, when tropical cyclogenesis and intensification is considered.

4. Interannual Fluctuations in Tropical Cyclone Numbers

4.1 El Nino/Southern Oscillation (ENSO) Effects

As noted in the previous section, interannual variability in any one of the cyclone regions (see Figure 2) is large, in spite of the overall global stability of cyclone frequency. In a number of ocean basins, a link between tropical cyclone activity in a particular season and phase of the Southern Oscillation has been drawn (e.g. Chan, 1985; Dong, 1988; Gray, 1984a,b; Shapiro, 1982a,b; Nicholls, 1989). In the Australian region a strong and significant positive correlation has been found between tropical cyclone frequency and the Southern Oscillation. That is, in years when the Southern Oscillation Index (SOI) is positive, corresponding to relatively warmer SSTs in the western Pacific and lower Darwin pressures (Nicholls, 1984), cyclone activity is increased. In years of large negative SOI when these trends are reversed (i.e. relatively cooler western Pacific SSTs and higher Darwin pressures), cyclone activity is suppressed in the Australian region (Nicholls, 1984; Allan, 1988).

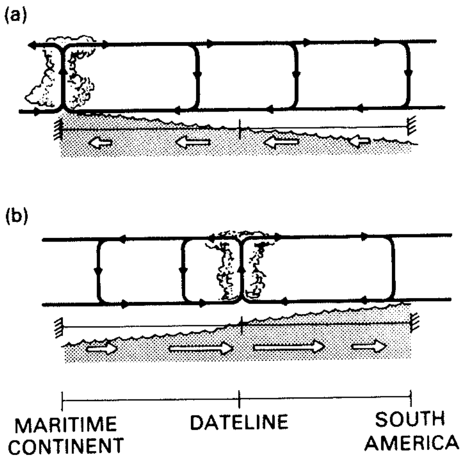

Figure 4: Schematics of a) non–El Nino and b) El Nino structure of the Walker circulation across the equatorial Pacific (Lau, 1982).

In these years, the centre of tropical storm activity moves further east and north, away from the eastern Australian coast (Revell and Goulter, 1986a,b; Nicholls, 1989).

Movement of the storm centre has been related to the changes observed in the tropical Walker circulation between ENSO and non–ENSO events (e.g. Allan, 1988). In a 'normal' year, or an anti–ENSO event (when the SOI is positive), there is large–scale convergence of moist maritime air in the North Australian–Indonesian region, resulting in a large region of deep tropical convection. This large–scale, deep convection and net upward motion provides an environment that is conducive to tropical cyclone development and maintenance (discussed further in the next section). However, in an El Nino year (when the SOI is negative), the SSTs in the central and eastern Pacific region are warmer than normal and the Walker circulation is observed to split, with the region of low–level convergence moving further towards the central Pacific and large–scale subsidence occurring in the North Australian region. This subsidence will tend to suppress tropical cyclone development in this region. Figure 4 is a schematic of the Walker circulation in ENSO and non–ENSO configurations. The regions of large–scale convergence and subsidence can be seen clearly.

Hastings (1990) compiled the relative distributions of tropical cyclone occurrence in the South Pacific region for five ENSO and three anti–ENSO seasons. These are illustrated in Figure 5 and clearly demonstrate the shift in storm activity away from the densely populated, eastern Australian coast in an ENSO year. This shift of storm activity results in decreased danger of tropical cyclone damage along the east Australian coast, while islands further out in the Pacific face an increased risk of cyclone strike.

ENSO events do not have a regular period, having been observed to occur at spacings of anything from two to ten years apart. The question of what triggers an ENSO event has yet to be answered, but the occurrence of westerly wind bursts in the equatorial western Pacific region just prior to the onset of an El Nino cycle has

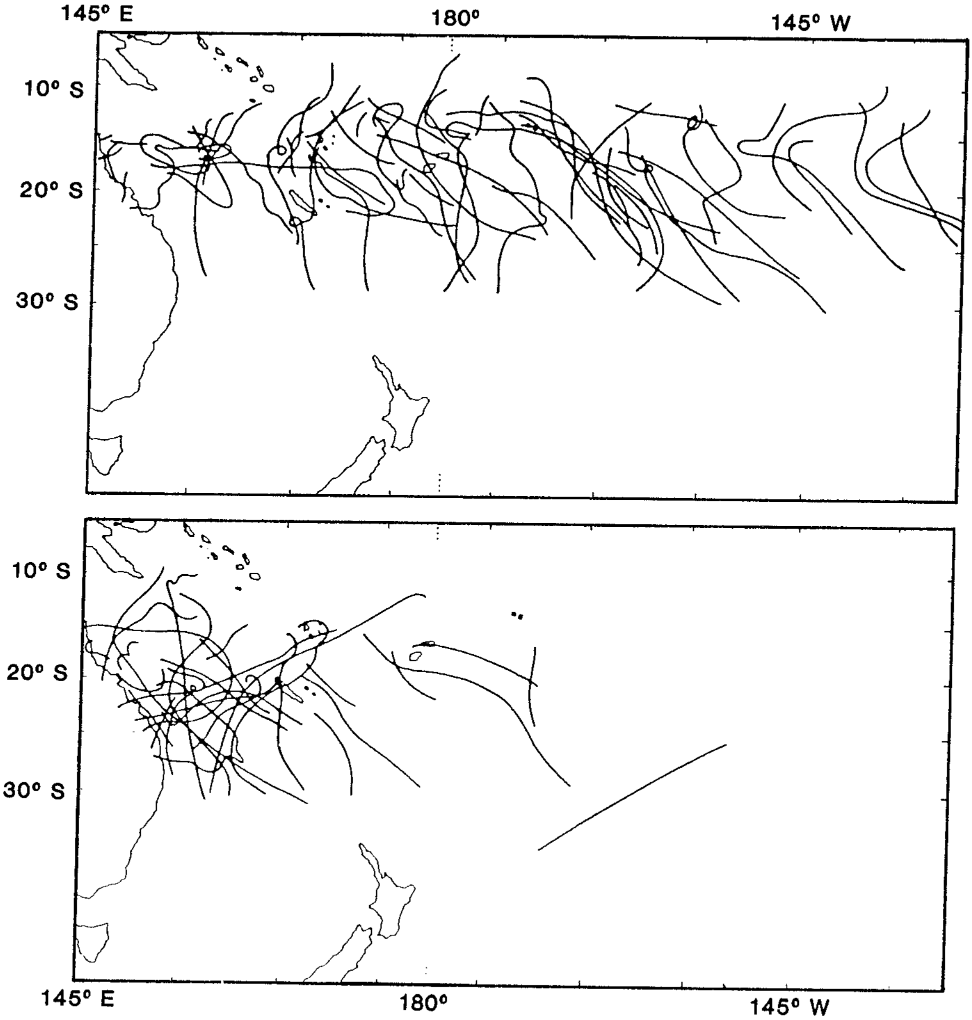

Figure 5: Distribution of tropical cyclones in the western South Pacific over a number of a) El Nino and b) non–El Nino years (Hastings, 1989).

been well documented (Nitta, 1989; Allan, 1988; Ramage, 1985; Lau, 1982;). The origin of these equatorial westerly wind bursts is still being debated, but their role in the ENSO onset is as follows: a westerly wind surge (either an increase in westerlies or a decrease of the trade easterlies) reduces the easterly surface stress on the ocean surface, causing a reduction in upwelling and an ultimate shallowing of the oceanic mixed layer. Also, it triggers an oceanic Kelvin wave which carries warmer water across to the East Pacific. Finally, Rossby waves propagating westward increase the warming of the central Pacific region.

Keen (1982) and Nitta (1989) have noted the presence of a pair of tropical cyclones in the central Pacific region approximately three months prior to the first observations of substantial warming of East Pacific SSTs during El Nino onset. This three month period corresponds well to the time required for an oceanic Kelvin wave to cross the Pacific basin. Chu (1988) shows evidence of mid–latitude forcing as a possible source of the westerly surges pointing out that these surges enhance the cyclonic vorticity at the equator and may be responsible for, rather than being caused by, the development of near–equatorial cyclone pairs observed by Keen (1982) and Nitta (1989). Shiying et al (1988) observed that an increase in the activity of near–equatorial tropical cyclones (not just cyclone pairs) over a season or more seems to coincide well with the development of an ENSO event. Ramage (1985) also proposed cyclones as a possible mechanism for triggering ENSO development.

If tropical cyclones do provide one of the triggers for ENSO development, then ENSO events will be under–predicted by current GCMs. However, Enfield (1989) has proposed that an alternative explanation of ENSO is that it results from an inherent ocean–atmosphere instability in the Pacific Ocean basin. The westerly surges would be interpreted as part of these changes in the ocean–atmosphere coupling, rather than a triggering mechanism. Debate on this cause–effect relationship continues, but obviously coupled ocean models need vast improvements before they can hope to model El Nino in either case.

In order to understand the effects of El Nino occurrence on tropical cyclones in a climate different to the present, the effects of this new climate on El Nino need to be addressed. If, for example, the frequency of ENSO events was to increase in a warmer, greenhouse climate, the long–term centre of tropical cyclone activity in the South Pacific would shift further from the Australian coast than it is at present. As discussed above for El Nino years in the current climate, this would result in a likely decrease in cyclone activity in the eastern and northern Australian regions, but the Pacific islands to the east probably would be adversely affected.

Currently, GCMs provide the best tool for interpreting the effects of increases in greenhouse gas concentrations on the large–scale weather patterns. Meehl (1990) and Washington (1990) discussed recent improvements in the NCAR Community Climate Model (CCM). Specifically, their atmospheric GCM has been coupled to a four–layer, 5° by 5° resolution ocean GCM. There is no Q–flux correction in this ocean model, but the drift away from energy neutral is only about 2 % of the interannual variation. Meehl (1990) notes that this coupled model produces interannual variations in the ocean and atmosphere that have some characteristics in common with observed large–scale features of ENSO and anti–ENSO events, mainly in the equatorial East Pacific. In spite of these encouraging signs, there are still many problems with the model 'ENSO events'. Generally, these problems can be traced back to a need for increased resolution. For example, the current oceanic resolution levels do not separate the Pacific and Indian ocean basins at the equator, thus preventing the observed warm pool from forming in the equatorial western Pacific. Also, the oceanic mixed layer is assumed to be a constant 50m deep and the next layer below this is 450m deep. This configuration can result in the mixed layer becoming too cool due to upwelling when the surface wind stress is high. The atmospheric GCM requires increases in resolution also, although these have currently been traded off against the need for a more realistic ocean representation. The current resolution limitations are tied to computational power and costs, since the spin–up time for the four–layer ocean model is on the order of decades before it can even be coupled to the atmospheric GCM. Hence, these resolution problems are likely to continue into the foreseeable future.

4.2 Effects of the Quasi–Biennial Oscillation of the Stratospheric Winds (QBO) on Cyclone Frequency

Another large–scale feature varying on a similar timescale to ENSO is the Quasi–Biennial Oscillation (QBO) of the stratospheric zonal winds.

Gray (1984a,b; 1988) has related tropical cyclone activity in the western Atlantic and western North Pacific to the phases of the QBO (Figure 6). He finds that for a westerly QBO, there are almost twice as many hurricane days in the west Atlantic than in an easterly phase and these storms are both stronger and longer lasting. In the western North Pacific this relationship is not observed to be as strong and in contrast to the Atlantic, Gray (1988) finds that cyclone activity equatorwards of 15° is enhanced in an easterly phase of the QBO. He attributes the relationship between cyclone activity and stratospheric wind direction to the perceived requirement that a tropical cyclone have small environmental vertical wind shear in its core. The QBO phases observed to correlate with enhanced cyclone activity in each region have reduced mean local vertical wind shear in the upper troposphere and lower stratosphere. This produces an environment more conducive to tropical cyclone formation.

These stratospheric wind oscillations are evident in the South Pacific and Indian oceans also, but to this author's knowledge, similar correlations for the Australian/Southwest Pacific region have not been investigated. It is possible that these effects may also be present there.

Maruyama and Tsuneoka (1988) presented evidence that a recent cycle of the QBO had been modified by the coincident occurrence of an ENSO event. The potential for modification of the QBO cycles by ENSO makes the problem of studying the interannual variability of the large–scale flows in which a cyclone will form even more complex.

At this stage, reports of validation of the GCMs used for climate scenarios have not included reference to the phases of the stratospheric winds modelled, possibly since these levels (around 50mb) occur at the top of the model domain and so may be affected by the model boundary conditions.

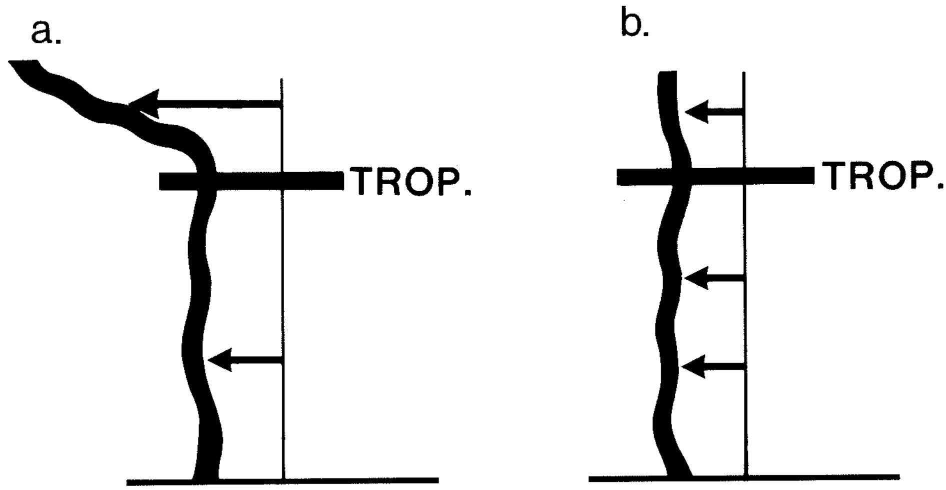

Figure 6: Idealisation of the vertical profile of the zonal wind in the tropics for a) east and b) west phases of the QBO (Gray, 1988).

5. Tropical Cyclone Genesis and Maintenance

5.1 Observed Necessary Conditions for Tropical Cyclogenesis

As noted in Section 3, a number of thermodynamic and dynamic conditions have been observed to be necessary for tropical cyclone formation.

Gray (1975) developed a seasonal genesis parameter (SGP) that incorporated the factors he had determined to be most important to tropical cyclogenesis. This parameter is defined as follows:

$$\begin{aligned}\text{SGP} &= (\text{oceanic energy}) \times (\text{vertical wind shear}) \\ &\quad \times (\text{humidity parameter}) \times (\text{absolute vorticity}) \\ &\quad \times (\text{inverse of atmospheric vertical stability})\end{aligned}$$

where detailed explanations of each of the terms can be found in either Gray (1975) or McBride (1981a). Each of these factors is considered in more detail below.

Thermodynamic factors that are seen to be favourable for tropical cyclone genesis include warm ($> 26.5^\circ$ C) ocean temperatures to a depth of approximately 60m, high humidities in the middle troposphere and possibly the potential for cumulus convection (i.e. net uplift in a conditionally unstable environment). This last condition will be considered further in Section 5.2 in relation to the CISK hypothesis for tropical cyclone development and maintenance.

As discussed by McBride (1981a), Frank (1987) and Holland et al (1988), these conditions are quite common in the tropical atmosphere, especially during the seasons when most cyclones are observed to occur. McBride (1981a) demonstrated this by calculating a 'thermal potential for genesis' (related to Gray's (1979) seasonal genesis potential, shown in Figure 7) for developing and non-developing composite

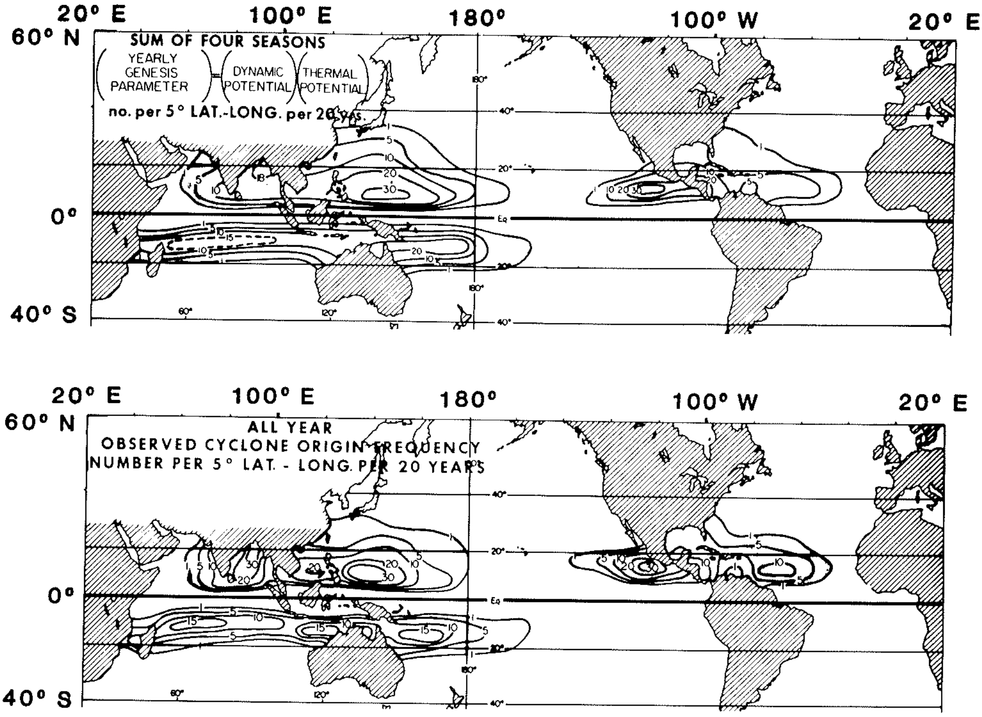

Figure 7: Global maps of a) Annual Genesis Potential (AGP) and b) observed tropical cyclone frequency, expressed in terms of occurrence over a twenty year period (Gray, 1975).

cloud clusters. The thermal potentials of both the developing and non–developing cloud clusters were essentially the same. Hence, the predicted increases in tropical SSTs and mid–tropospheric moisture may have little effect in the regions where cyclones currently form. However, these increases may extend the potential cyclogenesis regions, since the thermodynamic conditions listed above will be satisfied over a wider area. The greenhouse scenario considered here includes a warming in the tropics also, with the largest temperature increase foreshadowed to be in the middle to upper troposphere. This warming will act to stabilise the tropical atmosphere against convection. Taken in isolation, this temperature change may imply a decrease in tropical cyclone activity in the current genesis basins, due to suppression of large–scale convective activity. This idea will be extended in relation to theoretical ideas on tropical cyclogenesis presented in Section 5.2.

Since the thermodynamic requirements for genesis have been shown to be present throughout the cyclone season, the day to day variability of cyclone activity must be related to the dynamics of the genesis process (McBride, 1981a; Frank, 1987).

Gray (1975;1979) identified the dynamic factors contributing to tropical cyclogenesis as enhanced lower tropospheric absolute vorticity and small vertical wind shear over the centre of the incipient disturbance. McBride (1981a), McBride and

Zehr (1981) and Lee (1989) confirmed these for composite western North Pacific and North Atlantic storms.

Gray (1968) and Ramage (1974) have demonstrated that the most climatologically favourable area for tropical cyclone genesis in any ocean basin is in the region just poleward of the equatorial monsoon trough. McBride and Keenan (1982) have verified this relationship for the Australian region. Their study of cyclone formation over a five year period showed that 97 % of the cyclones occurring in the Australian region over this time were situated on the gradient–level (about 850mb), monsoon shear line at the time of classification to tropical cyclone intensity. Along the monsoon trough the vertical wind shear is very small, although to both the north and south the vertical and horizontal wind shear increases. This then is a region of enhanced lower tropospheric absolute vorticity. Large anticyclonic vertical shear in all directions around the incipient disturbance also was identified by McBride and Zehr (1981) as a favourable condition for cyclogenesis in the western North Pacific and Atlantic regions. They developed a 'daily genesis potential' based on the difference between the vorticity at 900mb and 200mb. If this exceeds a critical value and the wind field surrounding the disturbance is relatively symmetric, the incipient disturbance can be expected to develop.

McBride (1986) noted that the dynamic constraints listed above also govern the development of monsoon depressions in the Australian region and hypothesized that the further development of an eye structure in a tropical cyclone was due to its oceanic environment, but that the larger scale circulation was formed primarily by an eddy import of angular momentum at the upper levels (Holland and Merrill, 1984). Possible dynamical triggering mechanisms for cyclogenesis in the Australian region are the equatorward protrusion of an upper level trough (McBride and Keenan, 1982) or cross–equatorial cold surges from the Northern Hemisphere (Love, 1985).

In summary, the large–scale factors observed to be associated with tropical cyclone development in the current climate are warm sea surface temperatures over a relatively deep oceanic mixed layer, enhanced levels of absolute vorticity in the lower troposphere, weak vertical wind shear directly over the developing disturbance and mean upward motion coupled with high mid–level humidities.

The scenario assumed here for changes to the climate due to increased levels of greenhouse gases includes warmer tropical SSTs and increased mid–tropospheric moisture, as well as warming throughout the tropical region, with a maximum warming in the mid–troposphere. On the basis of preliminary results from existing GCM simulations, no significant shift of the climatological position of the monsoon shear line is expected. Hence, even though the thermodynamic conditions for tropical cyclogenesis (warmer SSTs, high atmospheric moisture) are likely to be satisfied over a more extensive region of the globe than they presently are, the possible increase in the stability of the tropical atmosphere may actually decrease the potential for tropical cyclones to develop. The relative stability of the position of the monsoon shear line foreseen for this warmer climate, if correct, indicates that the genesis region for tropical cyclones in the Australian region is likely to remain unchanged. A more confident statement on genesis locations must await improved GCM simulations.

5.2 Theories for Tropical Cyclone Genesis and Intensification – CISK and the Carnot Engine

Currently, there are two prominent analytical theories attempting to explain the genesis and intensification of tropical cyclones: the CISK hypothesis and the more recent concept of a tropical cyclone as an atmospheric Carnot engine. An outline of each of these ideas follows.

Conditional Instability of the Second Kind (CISK) is an extension of the ideas relating to cumulus convection. The basic concept is that the cumulus scales and larger scales co-operate via a linear instability, rather than acting in opposition, to spin up an initially weak vortex. Charney and Eliassen (1964) and Ooyama (1964) proposed that an initially weak, cyclonic vortex (idealised as an Eliassen balanced vortex) would draw warm, moist air in to its centre through boundary layer Ekman pumping. The air would converge near the centre and be forced up. The moist air would condense and form convective clouds. The latent heat released in these clouds would warm the middle levels, thus making it easier for the boundary layer air to be pumped aloft and form more clouds, strengthening the secondary, in-up-and-out circulation and deepening the vortex and so finally spinning up the tangential winds. By conservation of angular momentum, transport of cyclonic angular momentum away from the vortex in the upper outflow regime would spin up the upper-level anticyclone. This theory relies on the existence of a low-level, weak vortex and observations (Gray, 1975; McBride, 1981b; Lee, 1989) show that cyclones form in regions of enhanced lower level cyclonic vorticity, so this would seem an acceptable assumption. In order to close the equations, the CISK theory relates the middle-level heating to the boundary layer vorticity. Since a zero mean flow is assumed, this theory is not robust in its present form when one considers a moving system. Various modifications to this closure have been proposed, such as wave-CISK, where the heating is related to the vertical motion at the top of the boundary layer as well as the vertical motion in the mid-levels. The wave-CISK closure does allow propagation and is commonly used as a model for mid-latitude squall lines. If one assumes that the mid-level heating is merely proportional to the vertical motion at the same level, the instability generated is just conditional instability of the first kind and the heating generates local convective clouds.

Charney and Eliassen (1964) and Ooyama (1964) claim that all of the moisture is transported in through the boundary layer air and that no additional moisture was obtained by local evaporation of ocean water. In contrast, Emmanuel (1986) considers that the energy driving the tropical cyclone comes from evaporation at the ocean's surface – that is, the moist static energy deriving from surface fluxes drives the system. He considers an initially weak, convergent cyclonic vortex over a sufficiently warm ocean (SST $> 26.5^\circ$ C). As the air spirals inward, gaining moisture by evaporation, its speed increases by conservation of angular momentum. Since the rate of evaporation may be assumed proportional to the square of the surface wind speed, the moisture content of the near-surface air is increasing as it approaches the vortex centre, and it may approach saturation. On this inflowing path the air is cooled due to evaporation of ocean water, but an approximate balance is maintained between this process and the warming due to decreasing pressure as the air approaches the vortex centre. Since the air cannot continue to the vortex centre it flows upward in the eye wall cloud (along a moist adiabat) always warmer than the surrounding air and so conserves its energy. In the cirrus outflow region, the air flows outward, cooling by radiation to space and eventually sinking down to the surface again, far from the vortex centre. It may be drawn towards the vortex centre once again and the cycle repeats. Since the sensible and latent heat is gained by the system at a higher temperature than it is lost at, there is net mechanical energy available from this cycle. For a steady–state hurricane, this mechanical energy will balance friction, but in a developing system this additional mechanical energy will also act to 'spin–up' the system.

In this model, the cyclone circulation is viewed as a closed loop with respect to the atmosphere. The only energy source is at the ocean surface. This condition may be too restrictive, since cyclones are commonly observed to interact with their environment. In fact, Frank (1977b) and McBride (1981b) observed horizontal eddy fluxes above the boundary layer of similar magnitude to the energy imported within the boundary layer in their composite studies.

Emmanuel (1986) hypothesized that the most important thermodynamic interaction is between the vortex and the ocean (surface heat flux) with the cumulus convection redistributing the heat acquired at the oceanic source upward and outward to the upper tropospheric sink. This forcing mechanism is different from the CISK theory which requires a conditionally unstable atmosphere and organised convection. Whether or not the tropical atmosphere must be conditionally unstable for a cyclone to develop is of interest in the light of the current predictions for a stabilising of the tropical environment in a warmer, 'greenhouse' climate.

Rotunno and Emmanuel (1987) developed a nonhydrostatic, axisymmetric numerical model to test these ideas. They initialised their model so that the thermodynamic profile was conditionally neutral to their model clouds, then imposed a weak vortex circulation and let the model evolve. For a sufficiently strong initial vortex their model evolved and strengthened the vortex circulation. If the initial vortex was too weak, the circulation was maintained, but changed little over a 150h integration time. These results demonstrate that the surface flux mechanism can cause an initial vortex of sufficient intensity to develop further.

If the lower boundary condition is altered such that the SST is below about 26.5° C the model vortex will not grow. Sundqvist (1970a,b) proposed that the primary inhibiting factor to development here was the reduced surface moisture flux, rather than the surface heat flux. Hence, an initially weak vortex in either a conditionally neutral (Rotunno and Emmanuel, 1987) or conditionally unstable (Ogura, 1964; Kurihara and Tuleya, 1974) environment can intensify and grow, but this growth is heavily dependent on the surface fluxes.

5.3 Observed Limits on Tropical Cyclone Intensification

A number of studies (e.g. Miller, 1958; Merrill, 1988; Emanuel, 1987, 1988) have been carried out to determine the relationship between the maximum intensity of a tropical cyclone and the effects of its environment.

Miller (1958) related the minimum probable central pressure for a hurricane to the SST of the water it was passing over. He states that the empirical relationship he has derived is merely a lower bound for central pressure and that environmental effects can be expected to determine whether or not this central pressure is ever achieved. He derives a composite 200mb pattern for an environment conducive to deepening and a composite for inhibited deepening. Miller concludes that in order to achieve maximum deepening, maximum outflow and minimum inflow near the storm centre at 200mb are required.

Emanuel (1987, 1988) relates the maximum intensity (either the pressure minimum or wind maximum) of a tropical cyclone to the difference between the SST and temperature at the tropopause. The requirement for an upper–tropospheric temperature relationship comes from the Carnot engine model of the tropical cyclone proposed by Emanuel (1986). Basically, Emanuel (1987, 1988) says that the difference between the temperature at which heat is added to the cyclone system (in the boundary layer) and the temperature at which heat is lost from the system (the upper troposphere) and the thermodynamic efficiency of the system combine to provide the amount of energy remaining to spin up the cyclone. It is this relationship that he exploits to derive climatological potential maximum intensity charts. According to his theory, substantial SST increases in the tropics (2.3° C to 4.8° C, predicted by Hansen et al, 1983) may lead to substantial potential maximum intensity increases. However, this conclusion relies on the validity of the Carnot engine model of a tropical cyclone (yet to be determined) and on ideal environmental conditions existing.

Merrill (1987) compares climatological SSTs against observed maximum relative wind of all storms occurring in the Atlantic ocean over the period 1974 to 1985. He considers observed relative intensity – that is, observed maximum wind minus the translational velocity of the storm, since many of the highest intensities over cold water were associated with fast moving storms. Merrill's data is plotted in Figure 8. A sharp rise in maximum relative intensity above about 26° C is observed. No storm was observed to have a maximum intensity above 75 ms^{-1}. Merrill (1988) discusses environmental influences on the maximum hurricane intensity achieved and notes that while SSTs provide an upper bound, adverse environmental effects such as strong vertical wind shear across the storm's centre may prevent a storm attaining its maximum potential intensity. He derives a schematic pattern of upper level winds analogous to that of Miller (1958) for deepening versus non–deepening storms.

Obviously there is some relationship between the intensity of tropical cyclones and the SSTs. It is a matter of tropical meteorological folklore that no disturbance reaches tropical cyclone intensity over water colder than 26.5° C. As discussed in Section 5.1, the thermodynamic conditions for tropical cyclogenesis are largely seasonal, rather than daily factors. However, the upper limit of observed storm intensity has been shown to increase monotonically with climatological SST (e.g. Miller, 1958; Merrill, 1987, 1988). It would seem that environmental effects act to limit the maximum intensity of most storms observed. Otherwise, storms of the intensity of supertyphoon Tip in the western North Pacific or tropical cyclone Tracy in the Australian region would be observed more often.

For the majority of Australian region tropical cyclones (and intense tropical cyclones), intensification is related to upper–level interactions (Bureau of Meteorology, 1978; Holland, 1984a,b; Holland and Merrill, 1984). In fact, most strong tropical cyclones in this region develop just equatorward of the westerly subtropical

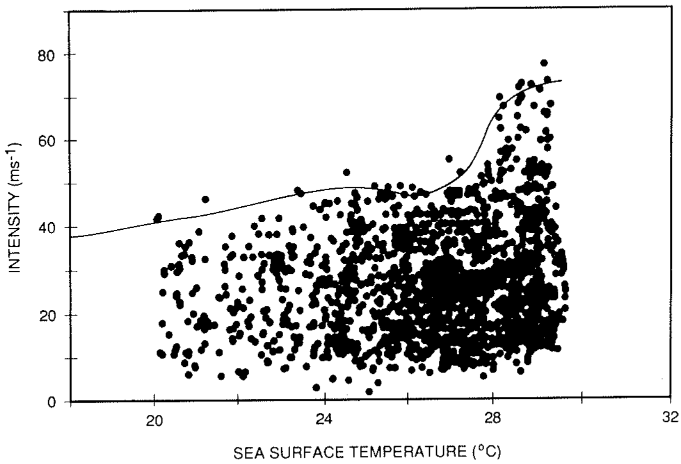

Figure 8: Empirically derived relationship between climatological SST and observed maximum hurricane intensity for Atlantic storms (Merrill, 1987).

jet. If a tropical cyclone moves under these strong westerlies, it will be 'sheared off', leaving a weak, shallow system (Holland, 1984c). The climatological studies upon which this data is based provide further evidence of the sensitivity to environmental conditions of the maximum intensity attained by a system.

6. Tropical Cyclone Structure

The structure of a tropical cyclone may be characterised in terms of three parameters: its intensity, size and strength. The intensity of a tropical cyclone is given by its maximum sustained wind (or equivalently, its minimum central pressure); the size of a system is a measure of the radial extent of, say, winds greater than 17ms^{-1}; cyclone strength gives an indication of the distribution of the strong winds – i.e. whether the system is compact, like cyclone Tracy (Director of Meteorology, 1977), or more dispersed, like cyclone Kerry (Lourensz, 1981). The potential effects of SSTs and the surrounding environment on the intensity of a system were discussed in Section 5.

A schematic representation of these three features is given in Figure 9 and radial profiles of the azimuthal winds of the two Australian cyclones, Tracy and Kerry, and the western North Pacific supertyphoon Tip, are shown in Figure 10. The potential variability of both cyclone size and strength, independent of intensity, is graphically illustrated by these azimuthal wind profiles. This result had been noted earlier by Merrill (1984) for Atlantic storms, although Weatherford and Gray (1988) found a weak relationship between intensity and outer wind strength in western North Pacific typhoons when they were stratified by eye size.

It is important to know the strength and size of a system making landfall, in addition to its intensity. All of these characteristics will affect the amount of wind damage experienced by the coastal communities of the region. For example, tropical cyclone Tracy was a very compact system, and a track deviation of only 100km or so

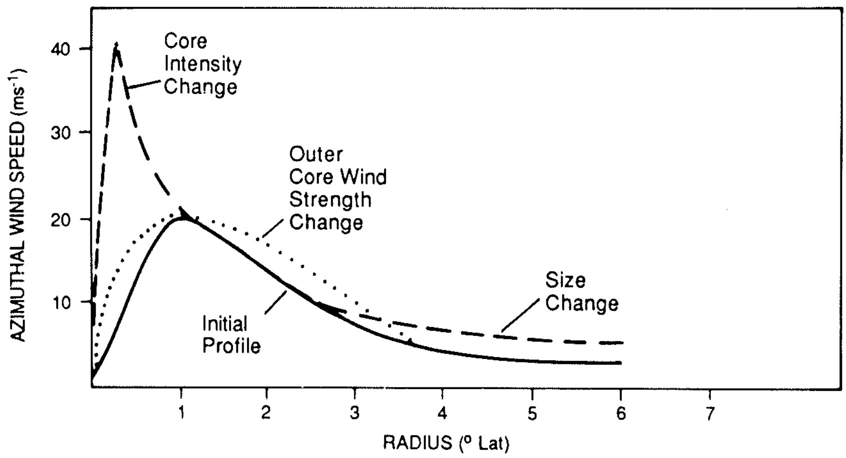

Figure 9: Schematic illustration of the concepts of *'intensity'*, *'strength'* and *'size'* of a tropical cyclone (Merrill, 1984).

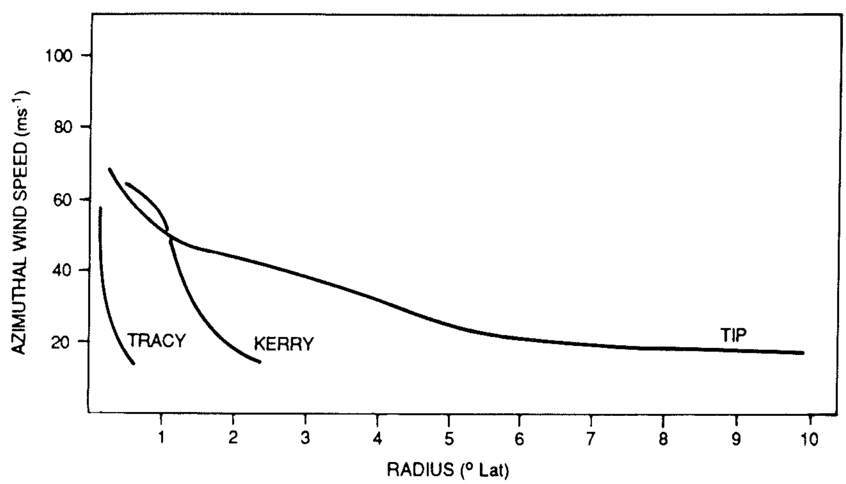

Figure 10: Radial profiles of the low-level azimuthal winds for tropical cyclones *Tracy* and *Kerry* and supertyphoon *Tip* (Holland and Merrill, 1984).

may have substantially decreased the impact of this storm on the people of Darwin. Unfortunately, there is very little understanding of the processes contributing to the relative spread of a tropical storm, so it is impossible at this stage to make sensible projections of changes in mean storm structure that may result from the greenhouse scenario assumed here.

7. Tropical Cyclone Motion

The concept of environmental advection as a contributing factor in tropical cyclone motion has long been accepted (e.g. James, 1950; Kasahara and Platzman, 1963; George and Gray, 1976; Bureau of Meteorology, 1978; Holland 1984a,b; Evans et al, 1990). This concept of large–scale advection is based on Helmholtz' theory of plane vortex interaction and really is only valid for the prediction of movement of vortices that are small compared to the scale of the flow field in which they are embedded. In practice meteorologists have found that, to first order, the mean horizontal wind averaged over the depth of the system (the 'deep–layer mean wind') provides a reasonable estimate of the motion of a tropical cyclone (e.g. Sanders and Burpee, 1968; Madala and Piacsek, 1975; Holland, 1983). However, there is a consistent discrepancy between the movement of a tropical cyclone and this derived 'steering flow' (e.g. Carr and Elsberry, 1990). Hence, other factors must be contributing to storm motion.

Many authors (e.g. Rossby, 1948; Adem, 1956; Anthes, 1982; Holland, 1983) have proposed that differential advection of Earth vorticity by the storm would provide another significant contribution to motion. Higher vorticity air would be advected from the poleward region to the west of the storm and lower vorticity to the east. These local maxima and minima of cyclonic vorticity will induce secondary circulations around themselves and will also be advected around by the storm. The gyres formed in this way will result in a net poleward and westward flow across the centre of the storm. This additional contribution to motion is known as the β–effect. Figure 11 is a schematic illustrating the proposed 'propagation' (i.e. motion in addition to large–scale advection). Evans (1990) and Evans et al (1990) extended this concept to include a contribution to propagation due to environmental vorticity gradient effects; that is, instead of propagating purely on the Earth's vorticity gradient, a cyclone in a dynamic environment propagates on the total background absolute vorticity gradient field. In this case, β in Figure 11 should be thought of as an absolute vorticity gradient and so need not remain oriented in the north–south direction. The potential complexity of the motion of a storm with this additional term is greatly increased. Shapiro and Ooyama (1990) note that, for timescales greater than a few days, this picture becomes even more complicated, due to non–linear feedbacks between the vortex and its environment.

Many other processes have been proposed as potentially contributing to tropical cyclone motion. Holland (1984b,c) noted that land effects to both the east and west of Australia modify the motion of storms in the region. Off the west coast, warmer continental air is advected to the poleward side of the storm, setting up an equatorward directed temperature gradient. From thermal wind considerations, this sets up a deep, easterly flow, which is modified by the β component of motion, causing the storm to track along the coast for long distances. Off the east coast, the

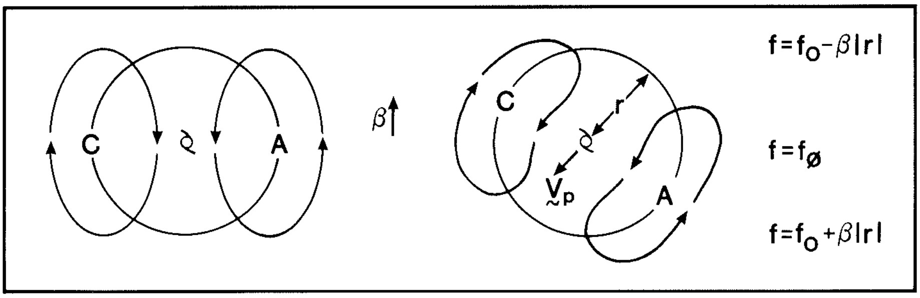

Figure 11: Illustration of the evolution of the so-called β–gyres (Evans, 1990).

strong, westerly, upper–level subtropical jet generally results in a mean eastward steering flow. This region is unique in that most of its storms move continuously eastward for a large part of their lives (Holland, 1984c). The presence of the Great Dividing Range has been noted (Holland, 1984a) to affect the low–level flow of a tropical cyclone near the coast and so may have some effect on the motion of the storm. However, the magnitude of this effect is not easily quantified.

Anthes and Hoke (1975) proposed that divergence effects may be important in modifying cyclone motion, however, Shapiro and Ooyama (1990) showed, in a shallow water equations model, that this is not the case. In their studies, Shapiro and Ooyama showed that divergence effects contributed a few kilometres track difference over a four day period, whereas varying the outer structure of the storm made a significant difference to the track. This difference comes from the variation in the β–effect due to the change in vorticity advection as well as through a change in the mutual distortion of storm and environment. DeMaria (1985, 1987) and Fiorino and Elsberry (1989) also demonstrated that changing the structure of the outer region of a storm (size, strength) had a large effect on storm displacement. However, when they doubled the intensity of a storm, keeping the outer regions constant, the track was unaffected. Baroclinic effects, such as the passage of Tropical Upper Tropospheric Troughs (TUTTs) or the intensification of the subtropical ridge, also have been proposed as affecting storm motion, but the mechanisms acting here are, as yet, not well understood.

Hence, any changes in the mean synoptic pattern may affect both the advective component of motion and the environmental β–effect directly. Changes in the environment also may affect the genesis locations (discussed in Section 5) and possibly indirectly affect the distribution of tropical cyclone storm sizes. Modifications to either the position or strength of the subtropical jet should have some effect on the movement and lifetime of tropical cyclones off the east coast of Australia. Warming of the continent may impact upon the magnitude of the thermal wind response off the Western Australian cost. However, at present the magnitudes of tropical cyclone responses to the various forcing changes mentioned here may not be readily quantified.

Features in the greenhouse scenario outlined in Section 2 which may impact on the motion of tropical cyclones are: warming over the land and changes in the structure of the large–scale wind field (including the positioning of the subtropical jet, for example). In Section 2 it was noted that considerable warming of the surface in the tropics (of the order of 2° C) is expected. The regional distribution of this warming is as yet, unclear, so nothing can be said about changes in motion due to variation in the land–sea temperature contrast. Hence, until more details can be gleaned from the GCMs (or other research efforts), one must assume that the land–sea interaction causing cyclones to track along the West Australian coast will remain much the same. The envisaged warmer waters farther south may allow cyclones to maintain their intensity for longer, however, as discussed in Section 5, this will depend on the co–existing atmospheric circulation also.

Our best estimate of the changes to important synoptic features, based on current GCM simulations, is that the monsoon trough may maintain its current mean position, perhaps strengthening a little, while the subtropical ridge may be displaced further poleward. This would result in a decreased environmental vorticity gradient and so would reduce, on average, the contribution to motion due to the environmental β–effect, although this will depend strongly on the positions of individual systems and the actual changes in intensity of the synoptic features. At least in the tropical regions, the environmental steering seems to be envisaged to remain largely unchanged, or perhaps increased slightly. Overall, the contributions to storm tracks discussed here can be expected to be largely unchanged, with the possible increase in environmental advection and slight decrease in the environmental β component. However, many more detailed regional analyses need to be made before even these broad tendencies can be stated with confidence. In addition, much work needs to be done in understanding the contributions to motion of the other processes just touched upon here.

8. Synthesis of Current Scenario

Various modelling and analogue studies predict a mean warming of the Earth's atmosphere due to the increasing levels of 'greenhouse gases' being released into the environment, largely due to human activity.

In order to attempt to quantify these effects and to determine the regional impacts of this warming, General Circulation Models (GCMs) have been developed. While many caveats still remain as to the accuracy of these models, they are the best tools available for this task at present. Hence, the common elements of six of these model simulations have been distilled into a working 'Enhanced greenhouse scenario' for this report. This scenario includes: decreasing surface temperature gradients from equator to poles, with the poles warming more than the equatorial regions in the lower levels; coincident with this atmospheric surface temperature rise, SSTs are expected to increase; in the tropics, warming is envisaged to increase from the surface to the middle troposphere, with a maximum increase of about 4° C near 400mb compared with only $2-3^\circ$ C at the surface; middle tropospheric moisture is also expected to increase; the lower stratosphere is envisaged to cool fairly uniformly from equator to pole. The envisaged temperature anomalies between the current climate and the climate predicted by this scenario are schematically represented in Figure 1. Three of the models have been studied in more detail and show no appreciable shift between the mean climatological position of the monsoon trough in their current climate and their 'greenhouse' climate. A slight strengthening of the monsoon may be expected and the subtropical ridge system is envisaged to be displaced further poleward in the Australian region.

Recent improvements to some of the models (including a coupled multi-level ocean GCM instead of a slab ocean mixed layer) indicate that the warming trends in the higher latitudes of the Southern Hemisphere may be lower than initially predicted, due to the moderating effects of the greater oceanic area in this hemisphere. However, this conjecture is based on a single model experiment with mean annual solar insolation. Inclusion of the seasonal cycle in these runs will better project the changes in the synoptic pattern likely to affect tropical cyclones, however, the models are unable to resolve tropical cyclones at this stage.

This report explores the potential changes in tropical cyclone behaviour in the Australian region as a result of the greenhouse scenario proposed here. A summary of the conclusions reached, or questions raised, follows.

(i) Climatology

Envisaged warming of SSTs and increasing moisture in the middle troposphere will extend the area of the Australian tropics for which the thermal conditions of tropical cyclone genesis are satisfied. This may allow for an increase in the number of tropical cyclones, but there are many caveats on this. These are dealt with in relation to interannual variability and genesis.

(ii) Genesis

The reasons for the observed lower limit on SSTs ($> 26.5^\circ$ C) before cyclone genesis occurs, are not well understood. This limit may relate to some mean property of the ocean–atmosphere system, rather than being an absolute condition. Hence, these thermal conditions may change in a warmer environment.

Even assuming that the thermal requirements for genesis remain as they are, it is dynamical factors that affect genesis on a daily basis. The projected stability of the current monsoon trough position in the warmer climate, if correct, is an indication that the genesis locations are likely to remain similar to those observed today. However, these mean genesis locations may be affected by an increase/decrease in the frequency of the El Nino/Southern Oscillation phenomenon.

(iii) Interannual Variability

The El Nino/Southern Oscillation phenomenon has been shown to modify tropical storm activity in the western South Pacific/Australian region. In a strong El Nino year, a modification of the Walker and Hadley circulations is observed. This results in an equatorward and eastward shift in the region of tropical convective activity, and hence in tropical storm genesis locations, compared to anti–El Nino years. Hence the centre of storm activity is much further out in the central Pacific region and the Australian region experiences a much quieter cyclone season than is usual. Incidence of tropical cyclones crossing the coast and moving inland is also necessarily reduced at these times. Much of the western and central Australian desert receives a significant fraction of its rainfall from dissipating tropical systems, so this shift in the seasonal pattern can severely reduce the amount of rain received in these regions, potentially causing much hardship.

At present, GCMs are just beginning to produce some of the synoptic features associated with an El Nino–type event. Hence, there is no clear indication as to whether or not the frequency of the El Nino cycle can be expected to change, so no prediction of a change in interannual variability of tropical cyclones in the Australian region is possible at this stage. The possible role of near–equatorial tropical cyclones in initiating El Nino events has recently been raised by a number of researchers. This only adds to the complexity of predicting the long–term trends.

(iv) Cyclone Intensity

The possibility of increasingly intense tropical cyclones in a warmer environment has been raised by a number of authors. This predicted increase is based upon the expected increase in tropical SSTs. Empirical studies have demonstrated a positive correlation between SST and the maximum observed cyclone intensity. The potential for more intense storms has been observed to increase with increasing SST. However, these studies also note that the majority of storms never achieve the potential maximum intensity possible, based upon these SST relationships. Environmental interactions are considered to dominate the intensification processes of a storm, limiting its intensity. Hence, merely warming the SSTs is not a sufficient condition to attaining more intense storms. The storm and its environment also must be working in concert for the maximum potential intensity of the storm to be reached. For example, if the subtropical jet were to move equatorward or strengthen, this could severely limit cyclone intensification off the east coast of Australia, whereas a poleward migration of the jet could allow more intense systems to develop. However, details of regional climate impacts are, at best, sketchy at this stage, and these are merely addition, possible, scenarios. It would seem that the potential will exist for more intense storms, but whether the environment will be favourable also is an issue that is yet to be addressed.

(v) Structure

There is no observed relationship between intensity and tropical cyclone size and only a weak relationship between intensity and strength (when eye size is taken into account), so even if it is assumed that storms will become more intense, nothing can be deduced about their wind profiles. Since the wind distribution of a tropical cyclone (e.g. the radius to maximum winds) affects storm damage forecasts and motion forecasts, this also restricts our predictions in these areas.

(vi) Tropical Cyclone Motion

The movement of a tropical cyclone is governed predominantly by advection by the large-scale flow (steering). A secondary contribution is the cyclone-induced motion (or propagation) on the environmental vorticity gradients. Other known contributions to motion come from topography, vertical wind shear and storm structure.

Any change in the large-scale wind pattern will affect the preferred cyclone paths, or cyclone 'corridors', currently observed. This may impact on rainfall, storm surge and wind damage statistics. Current indications are that the barotropic processes contributing to motion will only be slightly modified, with a potential increase in steering due to strengthening of the monsoon trough and perhaps a small decrease in propagation due to a slackening of the environmental vorticity gradients. However, this last point is very tenuous and must be regarded with some scepticism. Baroclinic effects have not been considered in much detail yet. If the environmental conditions are amenable, it is possible that warmer SSTs may result in tropical cyclones maintaining their intensity farther south than at present. However, more investigation is required to support this conjecture.

9. Synopsis

Overall then, if the current GCM simulations are to be believed, preferred cyclogenesis locations could remain unchanged. The potential maximum intensity of storms should increase, with the caveat that environmental effects are ultimately the limiting factor on actual cyclone intensity and motion will remain largely unchanged. At this stage, there is insufficient information available to enable more than speculation on future trends in the effects of ENSO and on the motion and structure effects of warm, continental air on cyclones, especially relevant to the West Australian region. Changes in the amount of rainfall derived from dissipating tropical systems are also unclear. Further emphasis needs to be placed on developing more complete regional scenarios for the 'greenhouse climate'. A better understanding of tropical cyclone genesis and intensification mechanisms and cyclone interaction with the larger scales are also high priorities.

In view of the present coarse resolution of the GCMs, it is the intention of this group to explore the sensitivity of tropical cyclone behaviour to various, postulated environmental changes using high resolution, limited area models.

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