On the dynamics of tropical cyclone structural changes

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SUMMARY

Tropical cyclone structural change is separated into three modes: intensity, strength and size. The possible physical mechanisms behind these three modes are examined using observations in the Australian/south-west Pacific region and an axisymmetric diagnostic model.

We propose that, though dependent on moist convection and other internal processes, the ultimate intensity, strength or size of a cyclone is regulated by interactions with its environment. Possible mechanisms whereby these interactions occur are described.

1. INTRODUCTION

Most of the work on tropical cyclone development in past years has been concerned with ‘genesis’ and ‘intensification’, even though no precise definition exists for either term. The result has been confusion in interpreting some aspects of cyclone structural changes, and the neglect of other aspects altogether. More precise and comprehensive interpretations are possible if we separate tropical cyclone structural changes into three modes: intensity, strength, and size (Merrill and Gray 1983). These are illustrated in Fig. 1 as changes from an initial tangential wind profile. ‘Intensity’ is defined by the maximum wind, or by the central pressure if no maximum wind data are available, and has received the most attention in the literature. ‘Strength’ is defined by the average relative angular momentum of the low-level inner circulation (inside 300 km radius). ‘Size’ is defined by the axisymmetric extent of gale force winds, or by the radius of the outermost closed isobar.

An extensive discussion on the need for an explicit consideration of strength and size in addition to intensity in describing tropical cyclones may be found in Merrill and Gray. The essence of this discussion is illustrated by the three radial profiles of azimuthal wind in Fig. 2. The Tracy and Kerry profiles were derived from data in Holland (1980) and archived synoptic analyses, and the Tip profile was derived from data in Dunnavan

![Figure 1](image)

Figure 1. A schematic of the effects of intensity, strength and size change on the radial profile of azimuthal winds in a tropical cyclone.

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and Diercks (1980). Even though their intensities were similar, these were obviously very different systems. Their approximate strengths and sizes were respectively: Tracy, $8 \times 10^5 \text{m}^2\text{s}^{-1}$ and 50 km; Kerry, $4 \times 10^6 \text{m}^2\text{s}^{-1}$ and 300 km; and Tip, $7 \times 10^6 \text{m}^2\text{s}^{-1}$ and 800 km. Thus, hurricane Tracy was an order of magnitude weaker and smaller than typhoon Tip. Indeed, the greater than gale force winds of hurricane Tracy could have been accommodated within the eye of hurricane Kerry!

Though the three modes are partially related, the correlations are small. Merrill (1982) has shown that there is only a weak correlation between intensity and size, and Weatherford and Gray (personal communication, 1984) have shown that strength and intensity changes often occur independently of one another. Intensity increase is often associated with a contraction of the radius of maximum winds; the maximum wind then increases by quasi-conservation of angular momentum as the inflow penetrates closer to the centre. The area and angular momentum requirements are quite small, however, so that pure intensity change can easily be accommodated by a slight internal rearrangement of the angular momentum fields. In contrast, an increase in strength or size requires a substantial import of angular momentum (see Holland 1983a). Many previous studies of environmental effects on tropical cyclones, particularly those involving angular momentum budgets, are therefore describing size or strength changes. As shown later, intensity is only indirectly related to outer region low-level angular momentum imports.

Confusion has also arisen concerning the term 'genesis'. Some take it to mean the initial appearance of a low-level vortex, while others see it as the development of winds of a certain speed. Much of the confusion has arisen because the progression varies from region to region; it means little to talk of origin of a closed low in a monsoon trough, while such an event is a milestone in a trade-wind environment. We prefer to define 'genesis' as the appearance of sustained low-level winds of $17 \text{m s}^{-1}$ or more, and use 'formation' to describe the process leading up to genesis.

In this paper we describe the controls on intensity, strength and size. In doing this we first compare axisymmetric observations from Australian/south-west Pacific composite cyclones with various diagnostic solutions of Eliassen's balanced vortex equations (appendix). We then examine the asymmetric dynamics of the cyclone/environment interaction. Finally, we present a discussion on the manner in which a cyclone forms, intensifies, strengthens and changes size.
2. **Axisymmetric vortex dynamics**

Axisymmetric cross-sections of radial and azimuthal winds for two composite tropical cyclones are shown in Figs. 3 and 4. These were derived by compositing observations around a sufficient number of similar tropical cyclones in the south-west Pacific region to produce quantitative analyses; AUS09 represents the intensifying tropical storm stage and AUS06 the intensifying hurricane stage of these systems. (Complete details may be found in Holland 1983b.) Note from the azimuthal wind cross-sections in Fig. 3 that these cyclones also grew and strengthened during the transition from tropical storm to hurricane intensity.

![Diagram showing axisymmetric cross-sections of azimuthal winds for two composite tropical cyclones AUS09 and AUS06.](image)

**Figure 3.** Axisymmetric, vertical cross-section of azimuthal winds (m s$^{-1}$) for the composite intensifying tropical storm (AUS09) and intensifying hurricane (AUS06) (after Holland 1983b).
Discussion will concentrate on the ways in which the distinctive features of the radial wind cross-section in Fig. 4 can be established. Specifically, we shall examine the extended outflow regime near 200 mb, the secondary inflow maximum near 400 mb and the two low-level inflow maxima.

These are commonly observed features of tropical cyclones throughout the world (e.g. Gray 1979, 1981). Their importance in determining the observed structural features in Fig. 3 may be summarized as follows. The primary low-level inflow maximum inside 6° latitude radius provides an import of angular momentum to offset the surface frictional

![Figure 4. Axisymmetric vertical cross-section of radial winds (m s$^{-1}$) for the composite intensifying tropical storm (AUS09) and intensifying hurricane (AUS06) (after Holland 1983b).]
dissipation and to strengthen the cyclone (Holland 1983a), together with a convergence of moisture to the core region convection (Ooyama 1964; Charney and Eliassen 1964). The deep inflow regime helps to maintain a strong, weakly sheared cyclonic wind regime in the middle troposphere. The secondary inflow maximum immediately below the upper tropospheric outflow ensures a sharp vertical wind shear from cyclonic to anticyclonic winds, which is largely in balance with a warm core maximum near 300 mb (Fig. 5.14 of Holland 1983b). The extended upper tropospheric outflow provides a removal of high potential temperature air to large radii, where return subsidence can be compensated by radiational or other cooling (Riehl 1954). Without such an extended outflow, return subsidence in the vicinity of the cyclone cores will tend to suppress convective activity; and, by adiabatic warming, weaken the horizontal temperature, and surface pressure, gradients across the maximum wind region. Finally, the low-level environmental inflow provides the substantial import of angular momentum required to increase the cyclone size (e.g. Merrill 1982).

(a) Background

The manner in which a cyclone will respond to a given environmental or internal forcing is dictated by its basic structure, with the dominant features being the ratio of inertial to static stability and, to a lesser extent, the baroclinicity (see Eqs. (A13), (A14) and (A15) of the appendix for definitions of these terms).

Physically, in an axisymmetric system, parcels leaving a constant forcing region are able to move horizontally or vertically depending on the inertial and static stabilities. For horizontal motion in an inertially stable region the radial gradient of angular momentum will cause an acceleration in the azimuthal wind speed, a change in the cyclostrophic and Coriolis forces, and a resistance to further horizontal motion (we neglect for simplicity the transient features of the geostrophic adjustment problem). The vertical density stratification in a statically stable atmosphere will similarly resist vertical motion. Mass continuity, however, requires that a circulation be complete. Then, as shown in Fig. 5, the scale of the induced circulation is determined by the relative effects of the inertial and static stabilities. For a relatively weak inertial stability, a long horizontal circulation will occur and the effects will be felt some distance from the forcing regions. A relatively strong inertial stability, however, will constrain the circulation to the near vicinity of the forcing. A similar result to that shown in Fig. 5 will occur for a vertical forcing, such as a heat source. As has been shown by Eliassen (1951), baroclinicity will also tilt the circulation axes in Fig. 5 as parcels tend to follow the sloping isentropic surfaces. And boundary conditions will further distort the circulations, as may be seen in work of Willoughby (1979), Schubert and Hack (1982) and Shapiro and Willoughby (1982).

A semi-quantitative indication of the likely tropical cyclone response to various forcing modes may be obtained from diagnostic solutions of Eliassen's balanced vortex equations, as described in the appendix. These equations form the basis of early balanced hurricane models (Ooyama 1969; Sundqvist 1970) and have been used by Willoughby (1979), Challa and Pfeffer (1980), Smith (1981), Schubert and Hack (1982) and Shapiro and Willoughby (1982) to examine secondary circulations and balanced response to sources of heat and momentum. We say semi-quantitative because the results are strictly applicable only to a balanced, perfectly axisymmetric system. Willoughby showed that the lower-level inner circulation is axisymmetric to a first approximation. But the upper levels, and in many cases the outer circulation, are often highly asymmetric. We shall consider these asymmetries presently, but at this stage will confine our discussion to the
response of an axisymmetric vortex to a core region heat source, and to outer region upper and lower momentum sources.

As an axisymmetric vortex we use the AUS06 data in Fig. 3; but with a core region added by presuming a maximum wind of 40 m s⁻¹ at 50 km radius and 850 mb; and with interpolation to a 30 mb by 10 km grid spacing using bicubic splines.

The resulting inertial and static stabilities (using Eqs. (A15) and (A13) of the appendix) are shown in Fig. 6. Note how there is a three orders-of-magnitude horizontal variation in the inertial stability, from a very stable core to a weakly stable outer circulation. Further, the lower troposphere is quite stable out to large radii, whereas the upper troposphere is only weakly stable. By comparison, the static stability is nearly constant below the stratospheric 'lid'. We shall demonstrate that the large horizontal variation in inertial stability provides the main constraint to the tropical cyclone response to different forcing mechanisms.

(b) The role of moist convection

The development and maintenance of the high energy core of a tropical cyclone is a very complex process, involving interactions between many scales. One mechanism, which has been summarized by Ooyama (1982) and forms the basis of extensive numerical modelling work, is the nonlinear cooperative interaction between the cyclone and cumulus scales. Basically, the clouds are arranged into consistent and organized patterns by the relative vorticity and inertial stability associated with the vortex. Entrainment into the clouds drives a secondary circulation with a deep inflow layer in the low to middle troposphere and a shallow, upper-level outflow. The clouds also support a local mass recycling which serves two functions: the first is to provide warming to keep the mass
field in balance with wind field evolving under the secondary circulation (Gray and Shea 1973); and the second is to enhance the oceanic evaporation and thus help the cyclone preserve energetic balance against the secondary circulation export of moist static energy (i.e. enthalpy plus potential) (Gray 1979).

The character and effectiveness of this cumulus–cyclone interaction also changes as the cyclone develops. During the initial stages the cyclonic vorticity, or inertial stability, is quite weak and any perturbations in the mass field are largely dispersed by gravity waves. Convective organization is poor, recycling occurs locally and only a very weak secondary circulation can be maintained on the cyclone scale. During intensification to the hurricane stage, the cyclonic core region vorticity increases and the convection becomes more organized, but the increasing inertial stability restricts the extent of the stronger secondary circulation. This restriction rapidly increases the efficiency of heating in the nascent eye region (Schubert and Hack 1982, 1983).

The secondary circulation, balanced vortex response of the AUS06 cyclone to an
imposed core region heating distributed as shown in Fig. 7, is contained in Fig. 8(a). (The streamfunction fields in Fig. 8(a), and all subsequent diagnostic results, are normalized to a maximum value of 10 to facilitate direct comparisons between different experiments.) In agreement with earlier work by Willoughby (1979) we see that two gyres are formed. Ascent occurs in the heated region with return subsidence in the eye and immediately outside; a deep inflow layer and a shallow, more intense outflow layer are also established. Note, however, the high inertial stability constraint on the outer circulation. Strong subsidence occurs just outside the heated region and the radial circulation becomes negligible beyond 2–3° latitude radius. Similar results have been obtained by Smith (1981).

(c) The role of the environment

It has long been known that tropical cyclones are limited to particular seasons and regions. Thus their existence, though made possible by the convective interactions described above, is largely regulated by the environment. These environmental controls fall loosely into two categories: necessary conditions which must be met for the cyclone to form or continue to develop; and sufficient conditions which force or enable such formation or development. The necessary conditions are well known (e.g. Gray 1968, 1975; Holland 1983b) and will not be considered here. Nor shall we consider the effects of sea surface temperature variations (see Holland 1983b for a complete discussion on this aspect). Rather, we examine the effects of dynamic environmental interactions in the upper and lower troposphere. Though these interactions are highly asymmetric and nonlinear (as we discuss in section 3), considerable physical insight can be obtained from a first consideration of the axisymmetric vortex responses.

The balanced vortex response of the AUS06 cyclone to an environmental imposition of an upper-level outflow is simulated by the upper momentum forcing profile in Fig. 7. The resulting secondary circulation is shown by the normalized streamfunction cross-section in Fig. 8(b). In this weak inertial stability region, a long shallow outflow layer is established, with return inflow in the stratosphere and between 400 and 600 mb. It is notable that this environmental forcing has a direct and substantial effect on the core region, even to the extent of generating lower stratospheric subsidence in the eye region.

By comparison, a low-level forcing, with the same distribution (Fig. 7), but of
Figure 8. Normalized streamfunction fields showing secondary circulation responses to the heat and momentum sources distributions in Fig. 7: (a) core region heating response; (b) upper-level environmental momentum forcing response; (c) lower-level environmental momentum forcing response.
opposite sign to generate an inflow, produces a quite different response. As may be seen in Fig. 8(c) the high inertial stability restricts the inward extent of the secondary circulation. Thus there is little direct core region effect from the environmental forcing. If moist processes were included in a conditionally unstable atmosphere the inward extent of this secondary circulation response would probably be further reduced.

\( (d) \) Synthesis

The radial wind fields resulting from each of the core region heating and upper and lower tropospheric environmental forcing are shown in Fig. 9, together with the response to a surface frictional dissipation which is zero at 880 mb and linearly increases to a surface value of \( F_s = -\rho r C_D |v_s| v_s \), where \( v_s \) is 0.8v at 850 mb, and, following Garratt (1977), \( C_D = (0.75 + 0.67 |v_s|) \times 10^{-3} \), here \( r \) denotes radial distance and \( \rho \) the density.

It is obvious from Fig. 9 that, taken alone, none of these mechanisms can adequately describe all the observed secondary circulations in Fig. 4. Further, as shown in Fig. 10(a), coupling the frictionally induced inflow and core region heating (along the lines of classical CIK theory) produces a deep inflow, with a low-level maximum, a shallow, constrained upper tropospheric outflow, and a secondary core region inflow maximum. This does not explain the observed upper tropospheric outflow (Fig. 4) and secondary 400 mb inflow maximum, nor the outer region low-level inflow maximum. These features are only obtained by including the environmental forcing, as may be seen in Fig. 10(b).

The following inferences may be drawn from these diagnosed axisymmetric circulation responses.

Enhanced core region convection may produce intensity or strength changes by an internal rearrangement of angular momentum but does not seem capable of affecting the size of the cyclone. As has been shown by Shapiro and Willoughby (1982) and Willoughby et al. (1982), this intensity change occurs as a nonlinear interaction in which a localized heat source (or convective band) outside the radius of maximum winds may result in the formation of a secondary eyewall and belt of maximum winds. This causes a dissipation of the primary eyewall and maximum wind region and temporarily decreases the cyclone intensity. The cyclone then re-intensifies as the new band of maximum winds contracts inwards by an acceleration of the winds just inside the maximum wind radius—a process described by Gray and Shea (1973).

However, as we have shown in Fig. 8(a), the heating in the high inertial stability core region also produces a detrimental, constrained circulation response. Strong return subsidence, with warming and convective suppression, occurs in the near vicinity. It is likely that the clear, doughnut-shaped region that surrounds many severe cyclones towards the end of a period of sustained intensification is due to this effect.

This constrained circulation can be counteracted by the extended outflow response to an imposed upper tropospheric environmental forcing. It is even possible that such a forcing could initiate an invigorated convective regime. Our diagnostic results in Fig. 8(b) also indicate that the environmental forcing may directly produce an intensification by driving eye region subsidence and an inflow maximum near 400 mb; both of these processes will enhance the warm core development.

Thus, enhanced core convection, coupled with upper tropospheric environmental forcing can provide sustained intensity, and possibly strength, changes. But size, and major strength changes require a substantial import of angular momentum, which can only be achieved by some form of lower tropospheric environmental forcing. (Note that imports by the frictionally induced inflow merely replace the frictional losses (Holland 1983a).) However, because of the high inertial stability, such low-level forcing does not
Figure 9. Normalized radial wind fields corresponding to the streamfunction fields in Fig. 8: (a) core region heating response; (b) upper-level environmental forcing response; (c) lower-level momentum forcing response. Also shown, (d), is the normalized radial wind field response to surface frictional dissipation.
directly affect the core region. Any intensity changes would occur as secondary responses by internal rearrangements following the initial outer circulation acceleration.

Recent numerical modelling results by Challa and Pfeffer (1980) provide some confirmation of our linear axisymmetric conclusions. They used Sundqvist's (1970) axisymmetric model (which is based on Eliassen's balanced equations) to examine tropical cyclone intensification under different imposed environmental forcing profiles. Though their discussion centres on eddy momentum transports, their modelling results are essentially a response to an axisymmetric forcing. An upper-level axisymmetric forcing produced an intensity change with very little size change. In contrast a low-level forcing produced a substantial size change, consequential intensity change, and development of a much stronger system. Compared to a reference run without forcing, Challa and Pfeffer found that the environmental forcing produced a more rapid intensification and more intense systems. They also noted that even with subcritical sea surface temperatures (and presumably weaker convection) intensification could be maintained by the environmental forcing.
3. ASYMMETRIC DYNAMICS

Since the observed secondary circulations in intensifying tropical cyclones require some form of environmental forcing, we proceed naturally to the question of how this can be achieved. Other important questions involve the validity of our linear, axisymmetric diagnosis, the effects of asymmetries, and the mechanisms whereby the core region and environment become coupled. A complete answer to these questions requires far more research than we have done in this study, together with better observations. However, our present knowledge does enable us to describe some of the more likely processes. To support our discussion, we shall use the wind fields for the composite AUS09 pre-hurricane tropical storm in Fig. 11. This is a composite of the developing tropical storm stage of those systems in the AUS06 composite described above. As may be seen in Holland (1983b) these are typical of the intensification stages of tropical cyclones throughout the Australian/south-west Pacific region.

(a) Low-level asymmetries

Let us first consider the low-level inflow into the outer regions of the intensifying storm (Fig. 9). We can see in Fig. 11 that this is maintained by a combined equatorial influx from the monsoonal westerlies and a subtropical influx from the trade wind easterlies.

Love (1982) has examined the role of the equatorial surge in the establishment and development of tropical cyclones. He proposed, and produced convincing supporting evidence, that this resulted from a surge originating in the winter hemisphere. The basic argument is that the winter hemisphere cold surge creates a high pressure region over the equator. In these low latitudes, an unbalanced down the pressure gradient surge is then developed which extends into the vicinity of the incipient depression. Subsequent pressure falls in the developing disturbance could also help maintain or enhance such a process.

The initiation of the subtropical trade wind surge has long been a forecasting rule for tropical depression intensification in the Australian region (e.g. Wilkie 1964). It is a quite transient effect associated with strong anticyclonogenesis poleward of the depression.

Both inflow mechanisms seem to be operating equally in the composite tropical storm of Fig. 11. But this is probably an artifact of the compositing process. It is more likely that the equatorial surge will be more important in low latitude systems and the trade wind surge will predominate at higher latitudes.

(b) The poleward outflow jet

The upper tropospheric analysis for the developing tropical storm in Fig. 11(b) shows that the principal outflow jet extends south-east of the centre and downstream of a strong confluenence between the storm outflow and subtropical westerlies. Such an outflow configuration has long been associated with tropical cyclone development in this region (McRae 1956) and has been linked with sustained intensification in other regions (Ramage 1959, 1974; Sadler 1976). It appears to be formed by a coupling between the cyclone outflow and passing disturbances in the subtropical westerlies.

A typical sequence of events is shown in Figs. 12 and 13, which contain the outflow layer variations during the main intensification and initial decay periods of hurricane Kerry, 1979 (minimum central pressure 954 mb, see Broadbridge (1981)). The observations in Fig. 12 and accompanying divergence patterns in Fig. 13 were provided by F. Lajioe (personal communication, 1982) and represent three short period composites of all available aircraft, satellite and conventional wind observations at a nominal 200 mb level.
At the commencement of hurricane intensification (Fig. 12(a)) the dominant feature is the strong poleward outflow into the divergent region ahead of an approaching westerly trough. Intensification continues (Fig. 12(b)) as the westerly trough amplifies, the subtropical jet stream moves past the hurricane, and the outflow jet strengthens. By 20 February (Figs. 12(c) and 13(c)), however, the upper trough has moved past the hurricane; convergent south-westerly flow has cut off the poleward outflow channel; hurricane Kerry begins to decay.

This decay is not typical of hurricanes in the south-west Pacific. As Holland (1983b) shows, the subtropical westerlies usually move over the hurricane and, literally, tear it apart. However, it is notable that with no further destructive, or enhancing environmental effects, hurricane Kerry spent a further 10 days over tropical waters and continued to decay gradually before crossing the Australian coast.
Figure 12. Outflow layer variations for hurricane Kerry during intensification: (a) February 16/00–17/00 GMT; (b) February 18/00–19/12 GMT; and (c) at the start of decay February 20/00–21/12 GMT (after Lajoie, personal communication 1982). Isotachs are in m s$^{-1}$. 
Figure 13. Divergence patterns ($10^{-5}$s$^{-1}$) derived from the hurricane Kerry composite outflow layer wind observations in Fig. 12 (after Lajoie, personal communication 1982).
(c) Equatorward outflow jet

In addition to the major poleward outflow channel, our Australian/south-west Pacific region composites also display a weaker, high-level equatorward outflow (Fig. 11(b)). This seems to be a consistent feature of most tropical cyclones in all ocean basins. Chen Lian-shou (personal communication, 1983) has shown that for the FGGE year at least 50% of all tropical cyclones had some form of equatorward outflow, and Sadler (1976) has indicated that a combined equatorward and poleward outflow jet is often associated with rapid intensification of north-west Pacific typhoons.

Since the inertial stability will be weaker on the equatorward side, this jet would be the expected result from a forcing by the hurricane convection combined with weak transient features on the equatorial easterly jet. That the outflow is strongest just below the tropopause and in the vicinity of deep moist convection also implies a large convective forcing component. However, this convective forcing appears to be more from the characteristic monsoonal feeder band in this region than from core region convection. In a trajectory analysis of a composite north-west Pacific typhoon, Lee (1982) suggested that the equatorial outflow jet, which was flowing outwards across absolute angular momentum surfaces, was maintained by an infusion of low-level cyclonic angular momentum by clouds along the major feeder bands.

Recent modelling results from DeMaria (personal communication, 1983) provides confirmation of these conclusions. DeMaria conducted two parallel experiments, one with, and one without, momentum transports in his cumulus parametrization scheme. With momentum transports a strong equatorward jet developed over the major feeder band. Without such transports a weaker and more constrained outflow resulted.

4. CONCLUDING DISCUSSION

We have made a simple examination of the ways in which a cyclone can interact with its environment by using observations and a linear diagnostic version of Eliassen's balanced vortex equations. This has lead us to the conclusions summarized in Fig. 14: (1) that upper tropospheric interactions can directly affect intensity change; (2) that

Figure 14. Schematic summary of the secondary circulations resulting from inner core convective heating and outer region momentum forcing, together with their effect on the tropical cyclone structure.
lower tropospheric interactions will directly produce a size change which, by subsequent nonlinear interactions, may indirectly affect the intensity, or strength of a tropical cyclone. Variations in inertial stability dominate in determining these different responses. In the low levels, the cyclone is very stable out to large radii, thus horizontal motion is constrained and large outer region accelerations may result from an imposed forcing. By comparison, the outflow layer has a quite low inertial stability, thus long radial trajectories are possible and an outer region forcing may directly affect the core region.

We have also shown that inner core convective heating may directly affect intensity and possibly strength change. Studies of intensity change by advanced numerical models have relied almost entirely on internal processes and have certainly produced intense hurricanes. As has been shown by Anthes (1972), these hurricane models develop an asymmetric outflow with strong jets by generating regions of inertial instability. Kitade (1980, Fig. 14) and Kurihara and Tuleya (1974, Fig. 10) show that large areas become unstable, and DeMaria (personal communication, 1983) finds that the local anticyclonic vorticity may approach twice the magnitude of the Coriolis vorticity. Yet, even though the outflow region is only weakly stable in nature, there has been no proof that large areas of instability develop: Alaka (1962) presented a detailed analysis of the outflow layer and could find at best very small, and presumably transient, instability regions; Black and Anthes (1971) found none. It is probable that the strong 200 mb outflow and 400 mb inflow near 2° latitude radius in Fig. 4 is partially due to convection and inertial instability in the unresolved core region (see Fig. 9(a)). But it is highly unlikely that the observed outer region outflow arises from core region convection or sustained inertial instability. The logical conclusion, then, is that without any environmental interaction, numerical models must take the alternative, and presumably more difficult, path of driving outflow jets by the development of large regions of instability.

Thus we suggest that cyclone/convective-scale interactions by the CISK process alone probably account for intensification only during the transition from tropical storm to minimal hurricane, when the eye is forming and convection is becoming well organized by the vortex. Initial development, intensification past the minimal hurricane stage, and rapid intensification, though dependent on moist convection, are probably controlled by some form of beneficial dynamical interaction between the cyclone and its environment.

We therefore propose that tropical cyclone development and subsequent intensification in the Australian/south-west Pacific region typically proceed as follows. Angular momentum transports by an initial surge in the monsoonal westerlies and/or trade wind easterlies generate a large, inertially stable monsoonal depression or shear zone. Provided this depression is in the benevolent environment described by Gray (1968), it can then loosely organize the fields of moist convection. This convection may in turn help intensify the depression by core heating and surface pressure falls (initially very weak), by an inward contraction of the maximum wind region, and by a vertical recycling of heat and momentum and enhanced oceanic evaporation. The more intense depression is then able to organize the convective fields better, and so on. Though this cooperative interaction is initially very inefficient, the efficiency increases exponentially as the system strengthens or intensifies. Thus, a strong monsoonal depression which has developed over northern Australia may rapidly intensify on moving over the ocean.

Of course, this development and intensification may be prematurely aborted by adverse environmental factors. The cyclone may move over land; or it may be sheared off by impinging strong upper tropospheric westerly winds (as Holland (1983b) has documented in the south-west Pacific), or by easterly winds (as occasionally happens in the northern Australian region).

An equatorward outflow channel will normally develop during the early formation
stage. This will be located just under the tropopause and will be maintained by vertical cyclonic momentum transport in the monsoonal feeder band.

Further intensification past the minimal hurricane stage in the Australian/south-west Pacific region then seems normally to occur as a result of a cooperative interaction between the cyclone and environment in the outflow layer. We suggest that this interaction starts during the tropical storm stage and proceeds as follows (see Fig. 15). A chance

![Diagram](image.png)

Figure 15. An illustration of the manner in which the subtropical westerlies and cyclone may interact to produce an extended poleward outflow channel.

...passage of a subtropical jet stream 600–800 km poleward of the cyclone and/or development of a westerly trough 800–1000 km upstream of the cyclone reduces the already low inertial stability in the outflow regime. It also generates a divergent area poleward and eastward of the cyclone centre. The initially constrained anticyclonic flow over the cyclone then turns and accelerates south-eastward to develop a long intense outflow channel. Core region inertial instability, created by active convection in the nascent eye region, with possible transient inertial instabilities associated with the passage of the jet stream, may aid this initial coupling.

Once established, this enhanced outflow may invigorate the core region convection, provide an inward contraction of the maximum wind region, and thus intensification. The outflow is also partially compensated by an inflow below the jet and in the lower stratosphere (Fig. 16). Differential angular momentum transports by the outflow and lower inflow generate a strong vertical wind shear between 400 and 150 mb and enhance the warm core development near 300 mb. The stratospheric inflow and concomitant eye region subsidence may also enhance this warm core development and provide further intensification (especially past the hurricane stage where wind to mass field adjustment becomes quite efficient).

The long outflow channel also allows compensating subsidence to occur over a large area. This reduces the debilitating subsidence heating and drying just outside the eyewall region that would result from the constrained circulation response to an increase in eyewall convection alone.

Of course the cooperation is almost certainly mutual. The cyclone outflow ahead of the westerly trough provides a strong warm advection. The temperature gradients
may be further enhanced by the frontogenetic effect of the confluent outflow and westerly wind regimes. Thus baroclinic, or barotropic processes may intensify the trough/subtropical jet couplet. It is even possible that the cyclone could initiate a perturbation in an initially zonal subtropical westerly flow.

This coupling is usually of short duration, say one to two days. It ceases when the trough moves past and cuts off the outflow channel to leave a slowly decaying cyclone, as happened with hurricane Kerry; or, more typically, when the westerlies move over the cyclone centre and strong shearing followed by rapid decay occurs (Holland 1983b).

In conclusion, this discussion has provided a physical description of the manner in which a tropical cyclone might interact with its environment to produce changes in intensity, strength and size. This description is consistent with previous research findings and with the documented observations of tropical cyclones in the Australian/south-west Pacific region. However, our present understanding of the actual underlying mechanisms can best be described as tentative. Further research towards delineating these mechanisms, and their relative importance, is needed and recommended.

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APPENDIX

The balanced vortex model

We begin with Eliassen (1951) balanced vortex equations in the form

\[ m^2/r^3 = \partial \Phi/\partial r \quad \partial \Phi/\partial p = -\alpha \] (A1, 2)

\[ \theta = (pa/R)(p_0/p)^k \quad dm^2/dt = 2m\chi \] (A3, 4)

\[ (1/r)(\partial u/\partial r) + \partial \omega/\partial p = 0 \quad c_p d(\ln \theta)/dt = Q/T \] (A5, 6)

where \( m = rv + fr^2/2 \) is the absolute angular momentum per unit mass, \( \Phi = \phi + f^2r^2/8 \), \( \chi \) and \( Q \) are the momentum and heat sources, \( d/\partial t = \partial/\partial t + u \partial/\partial r \) and \( \omega \partial/\partial p \) and the other terms have their usual meaning.

The hydrostatic equation (A2) and the gas law yield

\[-(\partial/\partial p)(\partial \Phi/\partial t) = \alpha \omega (\ln \theta)/\partial t \] (A7)

and the gradient wind equation (A1) yields

\[ \frac{\partial \omega}{\partial t} = -\frac{1}{r^3} \frac{\partial m^2}{\partial t} \] (A8)

Substituting Eqs. (A7), (A8) into Eqs. (A2), (A6) and (A4) we have

\[ -\frac{\partial \Phi}{\partial p} + \alpha \omega \frac{\partial}{\partial p} (\ln \theta) + \alpha u \frac{\partial}{\partial r} (\ln \theta) = \alpha Q/c_p T \] (A9)

\[ \frac{\partial \Phi}{\partial r} + \frac{\omega}{r^3} \frac{\partial m^2}{\partial p} + \frac{\mu}{r^2} \frac{\partial m^2}{\partial r} = 2m\chi/r^3. \] (A10)

We next define a streamfunction \( \psi \) such that

\[ u = \partial \psi/\partial p, \quad \omega = (1/r)\partial(\psi)/\partial r. \] (A11)

The continuity equation (A5) is then satisfied. Substituting Eq. (A11) into Eqs. (A9), (A10) and eliminating time derivatives then gives the diagnostic equation in \( \psi \)

\[ \frac{\partial}{\partial r} \left( A \frac{\partial \psi}{\partial r} + B \frac{\partial \psi}{\partial p} \right) + \frac{\partial}{\partial p} \left( B \frac{\partial \psi}{\partial r} + C \frac{\partial \psi}{\partial p} \right) = \frac{\partial E}{\partial r} + \frac{\partial F}{\partial p} \] (A12)

where

\[ A = -\alpha \partial (\ln \theta)/\partial p \] (static stability) \hspace{1cm} (A13)

\[ B = \alpha \partial (\ln \theta)/\partial r = -(1/r^2) \partial m^2/\partial p \] (baroclinicity) \hspace{1cm} (A14)

\[ C = (1/r^3) \partial m^2/\partial r = \{ f + (1/r) \partial(\psi)/\partial r \}(f + 2\alpha/r) \] (inertial stability) \hspace{1cm} (A15)

\[ E = \alpha Q/c_p T \] (thermal forcing) \hspace{1cm} (A16)

\[ F = 2m\chi/r^3 \] (momentum forcing). \hspace{1cm} (A17)

Equation (A12) was solved in finite difference form using successive over-relaxation on a 33 x 151 grid with spacing of 30 mb in the vertical and 10 km in the horizontal. Boundary conditions were \( \psi = 0 \) at the origin and upper and lower boundaries, and \( \partial \psi/\partial r \) constant at the outer boundary (which was sufficiently well removed to have no effect on the calculations).
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