Introduction

During the past fifty years, the problem on the formation and development of typhoons (hurricanes) has been studied by many investigators (Shaw, 1922; Sawyer, 1947; Haque, 1952; Syono, 1953; Yanai, 1961 a, 1961 b; Gentry, 1962; Shiroma and Sadler, 1965; Fett, 1966), notably Palmen (1948, 1956), Riehl (1948), Ramage (1959), Charney et. al., (1964), Gray (1968), and Sadler (1967 a, 1967 b). Unfortunately, opinions diversified to a certain extent (Yanai, 1964) and the lack of general agreement was evident (Gray, 1968). However, research workers (Chen, 1964; Simpson, 1970) have continued to contribute to enrich our knowledge in this problem, particularly in the field of numerical simulation and modelling (Kasahara, 1961; Kuo, 1965; Yamasaki, 1968 a, 1968 b; Ooyama, 1969 a, 1969 b; Rosenthal, 1969, 1970).

The primary objective of this lecture is to re-examine some of the findings of the pioneers in the light of recent numerical simulation experiments with a view to synthesizing a consistent unifying theory of typhoon development. No attempt is made to be exhaustive and a selective approach has been adopted in the present review. While I am aware that much remains unknown, the present state of our knowledge in typhoon development may be considered sufficiently rich for the derivation of certain forecasting procedures to benefit the practising forecasters.

Prerequisite and favourable conditions

In a well-known study, Palmen (1956) lists three prerequisite conditions for the development of intense tropical cyclones (including typhoons and hurricanes), namely:

1. "Sufficiently large sea or ocean areas with the temperature of the sea surface so high (above $26^\circ$ to $27^\circ$C) that an air mass, lifted from the lowest layers of the atmosphere (with about the same temperature as the sea) and expanded adiabatically with condensation, remains considerably warmer than the surrounding undisturbed atmosphere at least up to a level of about 12 km".

2. "The value of the Coriolis parameter larger than a certain minimum value, thus excluding a belt of the width of about 5 (to 8) deg lat on both sides of the Equator".

3. "Weak vertical wind shear in the basic current, thus limiting formation to latitudes far Equatorwords of the subtropical jet stream".

Riehl (1948, 1954) contributes two additional requirements which may be considered as favourable conditions, i.e.,
4. A pre-existing low-level disturbance (areas of bad weather and relatively low pressure).

5. Upper tropospheric outflow above the surface disturbance.

To this collection may be added the following:


(For convenience, they will be referred to as Conditions 1 to 6 hereafter in this lecture.)

Precursory signs

With the advent of the satellite era, cloud pictures received from orbiting weather satellites under operational conditions have proved to be a valuable source of real-time information for the recognition and diagnosis of precursory signs for tropical cyclone development in general. Our main interest lies in the identification and analysis of cloud clusters (Figure 1) over tropical seas or oceans, because they are characterized by the same scales of motion in space and time as tropical cyclones, of which typhoons (hurricanes) are the intense ones (Table 1).

In practice, it is convenient to use the scheme introduced by Williams (1970), which classifies cloud clusters mainly on the basis of their daily life cycles into five basic categories (Figure 2) as follows:

1. Developing cluster;

2. Dying cluster;

3. Developing-dying cluster;

4. Conservative cluster;

5. Pre-storm cluster - that eventually formed tropical storm or typhoon (hurricane).

There was evidence that these cloud clusters were produced and maintained by low-level frictionally forced convergence, which Charney and Eliassen (1964) call "conditional instability of the second kind", or CISK, in the zonal shearing environment north of the Equatorial trough (Gray, 1968). Recent observational analysis (Williams, 1970) lends further support to this contention.

Although vertical shears at the centres of all types of trade wind clusters are very small (within ± 8 knots) by mid-latitude standard, only about 13 per cent of a large sample of 1,257 clusters studied (Williams, 1970) belong to the pre-storm category. In general, pre-storm clusters are characterized by an easterly vertical wind shear at their centres, i.e. stronger easterlies aloft than near the surface (Figure 3). These findings are of paramount importance, because not only can they serve as definitive discriminators to differentiate pre-storm clusters from all other types, but they can also be used to determine when an ordinary cloud cluster has
become a pre-storm cluster as a useful precursory sign of tropical cyclone development, provided that the observation network is sufficiently dense.

Pre-storm clusters differ from other types in many respects. While significant differences in the low-level horizontal wind shears exist among all categories, pre-storm clusters exhibit greater low-level cyclonic vorticity in both the shear and curvature components but with a smaller shear/curvature ratio than all other types (Table 2).

Development theories and numerical simulation models

In an attempt to explain the formation and development of typhoons (hurricanes), many theories have been advanced by meteorologists. Riehl (1948) considered typhoon development as a progressive process of intensification of a migratory disturbance or wave embedded in the trade winds which moves under a favourable upper-tropospheric divergent environment. A recent study by Simpson (1970) on the influence of the stimulation of mass circulation reinforced Riehl's hypothesis (Figures 4 and 5). The analysis of Gray (1968) demonstrated that within 20° of the Equator, initial tropical storm formation occurs just poleward of the doldrum equatorial trough within a region of large low-level cyclonic horizontal shear (Figure 6) with minimum vertical wind shear. This covers development in a broad easterly basic current. However, there was evidence (Shiroma and Sadler, 1965; Sadler, 1967b) of development within a low-level monsoon trough which has a basic westerly current on the south side prior to intensification. This valuable finding has led Brody and Jarrell (1969) to the formulation of a practical method for predicting tropical cyclone genesis in the South China Sea.

In addition, Sadler (1967a) has analysed typhoon development, mainly in latitudes north of 20°N, resulting from downward tropospheric intensification of a pre-existing upper tropospheric trough.

Ramage (1959) advocated an energy dispersion mechanism (from an external source) to account for typhoon development. Energy of inertia-gravity waves generated by an intensifying upper trough or cyclone which shows a tendency of adjustment towards a balanced state is propagated from the source at group velocity and almost always travels much faster than the originating disturbance. From mid-latitude experience (Bjerknes, 1951), it may be inferred that in low latitudes, the divergence field cannot adjust rapidly enough to prevent deepening in the next downstream trough. Thus, a low-level disturbance located beneath or near the upper trough axis might deepen and develop into a typhoon or hurricane (Figure 7). In support of this hypothesis, Ramage lists the following:

"(a) Equatorward (poleward) of the subtropical ridge typhoon or hurricane development in a surface disturbance most often occurs beneath poleward (Equatorward) flow shortly after an upper trough or cyclone intensifies to the east (west);"

(b) Some surface depressions whose circulations extend weakly into the high troposphere or former upper cyclones which have become warm-cored may transform into typhoons or hurricanes if they are located downstream from an intensifying upper cyclone or trough;
(c) The most active typhoon-(hurricane-) generating area in the world, between 130°E and 150°E in the tropical North Pacific, is approximately one wavelength (about 24° of longitude) downstream from the most persistent and active upper trough in the tropics (now usually known as "Mid-Pacific Trough" or MPT after Ramage);

(d) Hurricane winds usually first appear in a storm's north-east quadrant where presumably vertical shear is least."

The early stage of typhoon development was investigated by Chen (1964) in relation to adjustment towards thermal wind balance at low latitudes in a conditionally unstable atmosphere, which is characterized by a greater environmental lapse-rate ($\gamma$) than the saturated adiabatic lapse-rate ($\gamma_m$), under the assumption of sufficient moisture supply. He defined the characteristic scale of such an adjustment process ($L_0$) as

$$L_0 = Rf^{-1} \sqrt{T(\gamma - \gamma_m)/2g}$$  \hspace{1cm} (1)

where $R$ = specific gas constant for dry air;

$f$ = coriolis parameter;

$g$ = acceleration of gravity;

$\gamma$ = environmental lapse-rate

$\gamma_m$ = saturated adiabatic lapse-rate;

$T$ = mean temperature of the air column under consideration. (Taken to be 260°C or -13°C for the tropical atmosphere).

This characteristic scale of the adjustment process is shown in Figure 8 as a function of latitude and the excess of environmental lapse-rate over the saturated adiabatic lapse-rate for a range of values appropriate for the mean tropical atmosphere. If initially the vorticity of the vertical shear in the wind field of an incipient disturbance is greater than the vorticity of its mean temperature field between the upper and lower levels (200 mb and 850 mb) and its horizontal scale ($R_0$) is less than the characteristic scale of adjustment, i.e.,

$$\nabla^2_h \left( \Psi_{T=250} - \Psi_{T=850} \right) > f^{-1} \nabla^2_h \left( \Phi_{T=250} - \Phi_{T=850} \right)$$  \hspace{1cm} (2)

and

$$R_0 < L_0$$  \hspace{1cm} (3)

where

$$\nabla^2_h = \left( \frac{\partial}{\partial x^2} + \frac{\partial}{\partial y^2} \right)$$

$\Psi_{T=250}$, $\Psi_{T=850}$ - stream function at 250 mb, 750 mb;
\[
\phi_{-250}, \phi_{+750} = \text{geopotential at 250 mb, 750 mb in geopotential metre (gpm)},
\]

then convective instability of inertia-gravity waves will occur. This may lead to rapid intensification of the incipient disturbance to a violent vortex in a relatively short time (about 24 hours) with high-level anticyclonic over low-level cyclonic circulation and a warm-core structure. These features are similar to those observed in typhoons and hurricanes. The computed development of a weak incipient disturbance (Figure 9) into a violent vortex is shown in Figure 10. It may be noted from Figure 10 that no significant vortex development occurs equatorward of 5°N. This is due to very small values of the Coriolis parameter near the Equator and thus substantiates Palmen's second specification (Condition 2).

In addition, the condition of \( R_o \ll L_o \) requires that
\[
\frac{\partial}{\partial t} \left[ \nabla^2_{h} (\Psi - \Psi_{750}) - f^{-1} \nabla^1_{h} (\phi_{-750} - \phi_{+750}) \right] > 0 \quad (4)
\]
which means that the rate of decrease of the vorticity of vertical shear in the wind field is less than that in the mean temperature (thickness) field. At this stage, it must be borne in mind that the cyclonic vorticity in the mean temperature field decreases very rapidly as the warm-core builds up. Consequently, upward motion within the domain of the incipient disturbance tends to amplify the initial imbalance of the thermal wind relationship, and thus a continual imbalance is maintained.

Conversely, if \( R_o \geq L_o \), we have
\[
\frac{\partial}{\partial t} \left[ \nabla^2_{h} (\Psi - \Psi_{750}) - f^{-1} \nabla^1_{h} (\phi_{-750} - \phi_{+750}) \right] < 0 \quad (5)
\]
This requires that the initial imbalance in the thermal wind relationship decreases with time and the system will soon attain thermal wind balance when the inertia-gravity waves are said to be neutral. Although the equilibrium may be disrupted by other factors as time goes on, it fails to trigger a chain reaction in the disturbance to bring about growth of the disturbance. These theoretical deductions are supported by an extensive discussion on the characteristics of the vertical motion within the hurricane presented by Gray (1966), who stated: "Cumulus-produced horizontal acceleration accounts for the imbalance in the thermal wind relationship. Tropical storm development is not viewed as being possible unless the cumulus-induced vertical momentum transfers act in a dominant way to oppose the thermal wind requirement and inhibits increase of vertical wind shear as baroclinicity increases. Vortex development thus requires a continual imbalance of pressure over wind acceleration".

The chain reaction of rapid development triggered by the instability of inertia-gravity waves as a result of adjustment towards thermal wind balance may be viewed as an internal energy dispersion mechanism of development in contrast to Ramage's hypothesis of external energy dispersion, because the adjustment process is accompanied by the generation of inertia-gravity waves to disperse the perturbation energy of the imbalance component from the source to other areas within the system or to the environment. This primarily covers cases of rapid development.
Now let us turn our attention to gradual development brought about by evolution process in which advection is predominant. Much mathematical experience on numerical simulation experiments (Kuo, 1965; Ooyama, 1969 a; Rosenthal, 1969) has been gained in parameterizing cumulus convection and heat fluxes from sea to air in the development of intense tropical cyclones. Meanwhile, numerical experiment in forecasting hurricane development with real data has achieved remarkable success in a case study (Miller, 1969).

The numerical simulation of the life cycle of tropical cyclones carried out by Ooyama (1969 a) uses the balance approximation and is primarily concerned with the development of an axi-symmetric, quasi-balanced vortex (Figure 11) in a stratified incompressible fluid formulated on the basis of "a co-operative process between the cumulus-scale moist convection and the cyclone-scale circulation. The sea-air exchange of heat and angular momentum was explicitly calculated. The results (Figures 12 and 13) show that the model is capable of simulating the typical life cycle of tropical cyclones with a remarkable degree of reality.

In this model, one conspicuous defect noted was a somewhat large radius of the eye (about 40 km), which tended to expand too rapidly during the mature stage. This was thought to be due mainly to the use of the balance approximation in the calculation of the boundary-layer inflow. With a view to removing this defect, Ooyama (1969 b) used the primitive equation of motion under the quasi-steady-state assumption to compute the wind in the boundary layer and obtained more realistic results (Figures 14, 15 and 16), namely, the radius of the eye is now 15 to 20 km and does not expand so rapidly as it did with the balanced model.

Test results on the response* of the model cyclone to changes in sea-surface temperature (Figure 17) show that both the intensity and size of the simulated cyclone are sensitive to changes in sea-surface temperature and the size of the warm sea area. This renders full support to Palmen's Condition (1).

The developments of the simulated tropical cyclones from weak initial vortices (Max $v_1 = 5 \text{ m sec}^{-1}$) of various sizes (radii of max $v_1 : 25 - 100 \text{ km}$) were also computed but the results (Figure 18) show that they take an usually long time to develop to typhoon (hurricane) intensity. This aspect was recently discussed by Ramage (1969), who stated:

"Although measurements are seldom made in weak disturbances, the time required for a 5 m sec$^{-1}$ circulation to develop into a typhoon is almost always less than 100 hr, compared to more than 500 hr (in the extreme case) for the model (and about 400 to 130 hr in broad general terms)."

In addition, Ramage also presented evidence (Figure 19) to mark the close resemblance of the model and nature during the later stages of typhoon (hurricane) development and during the mature stage, and certain useful statistics, which are recapitulated in Table 3.

*The response characteristics affecting the weakening of the simulated cyclone will be discussed in more detail in my subsequent lecture on "Typhoon dissipation".
Simulation experiments using the primitive equation of motion (Yamasaki, 1968 a, 1968 b; Rosenthal, 1969) produced results which are fairly consistent with those obtained by Ooyama (1969 a, 1969 b). Rosenthal's results are particularly good for illustration (Figures 20, 21 and 22).

A common feature in the primitive modelling experiments is that there is little change of central pressure or maximum surface wind in the simulated cyclone during the first three to four days. Intensification occurs in the next three days. In the latest experiment, which includes an explicit water-vapour cycle, Rosenthal (1970) reported that in the moist environment growth to the mature stage becomes more rapid and the total time taken is reduced to about four days, but the peak intensity is not strongly affected by the initial moisture content. But we must bear in mind that numerical simulation experiments performed so far are primarily concerned with typhoon development through evolution processes in which advection plays the major role. This takes four to eight days to produce a warm core with a 5 - 10°C temperature rise, whereas if adjustment processes (which involve energy dispersion) predominate, the time taken is reduced to less than one day (20 hr) to produce a (+9°C) warm core.

It is a pity that the characteristics of the structure of the simulated cyclone just prior to marked intensification were not reported by investigators in numerical modelling with the primitive equation. I opine that the computed structure 12 - 24 hours before rapid intensification can lead us to a better understanding on the development mechanism than illustrations of the structure at peak intensity and other times. The evidence presented in this lecture indicates that upper tropospheric outflow is an accompanying feature or a consequence of adaptive development and a precursor or cause of advective development. Thus this is in agreement with Priestley's emphasis (1956) of almost simultaneous occurrence of upper divergence and lower convergence.

With the foregoing background information, we are now in a position to synthesize all the evidence into a unifying theory of typhoon (hurricane) development. The processes involved in the various phases may be represented schematically in Figure 23.
NECESSARY DATA AND TOOLS

For the prediction of typhoon development we need the following:

1. Charts
   (a) Sea-surface temperature charts
   (b) 1000 mb charts
   (c) 850 mb contour charts
   (d) 850 mb streamline charts with isotachs
   (e) 200 mb contour charts
   (f) 200 mb streamline charts with isotachs
   (g) 200-850 mb thickness charts
   (h) 200-850 mb vertical shear charts with shear streamlines and isotachs

2. Satellite photographs

3. T - z gram plots of radiosonde ascents from stations or weather ships within the domain of the incipient disturbance

4. Tables 2, 4, 5 and 6 and Figures 3, 4, 7, 8 and 23 for reference

5. Grids
   (a) Grid for computation of vorticity from contours or thickness lines
   (b) Grid for computation of shear component vorticity from streamlines and isotachs
   (c) Scale for computation of curvature component vorticity from streamlines and isotachs.
FORECASTING PROCEDURES

For practical purposes, the following procedures are proposed for forecasting typhoon development. The suggested criteria will be refined for specific regions in due course.

1. Draw isotherms of sea-surface temperature and mark off warm sea areas (WSA) where sea-temp. $\geq 27^\circ$C.

2. Look for presence of low-level disturbances within WSA (e.g. zones of strong cyclonic horizontal shear north of Gray's equatorial trough below 850 mb, or near the shear line at 850 mb in Sadler's monsoon trough, easterly waves, vortices and discrete cloud clusters if latest satellite pictures are available).

3. Determine the size of the incipient disturbance ($R_o$) as follows:

   (a) Take the shortest radius of the outermost closed isobar drawn to the nearest mb (in deg lat); (1 deg lat. = 60 n mi);

   (b) If the incipient disturbance is also characterized by a cloud cluster, take half the estimated mean diameter of the cloud mass after exclusion of the cirrus canopy;

   (c) If in doubt use the arithmetic mean of (a) and (b).

4. Use any $T - \zeta$ gram available within the circulation of the incipient disturbance to estimate $\left( \gamma - \gamma_m \right)$ and check with the aid of Figure 8 to ensure that $R_o < L_o$. If such $T - \zeta$ grams are unavailable, estimate $\left( \gamma - \gamma_m \right)$ from Figure 8 for $R_o < L_o$. If the value of $\left( \gamma - \gamma_m \right)$ so estimated is reasonably consistent with latest past observations, proceed to 5. (In case of inconsistency, either stop or treat forecast with low confidence at the discretion of the forecaster).

5. At the centre of each low-level disturbance (or any you are interested in),

   (a) Compute* vorticity (units $10^{-5} \text{ sec}^{-1}$) in the 200 - 850 mb thickness field $\left( \nabla^2 \gamma^2 \zeta \right)$;

   (b) Compute vorticity (units: $10^{-5} \text{ sec}^{-1}$) in the 200 - 850 mb vertical wind-shear field $\left( \nabla^2 \gamma \|^2 \right)$;

   (c) Compute $D = \left( \nabla^2 \gamma^2 \zeta - \nabla^2 \gamma \|^2 \right)$ in $10^{-5} \text{ sec}^{-1}$.

*Procedures for the computation of vorticity will be dealt with in the practical sessions with the aid of grids.
Since development depends on many factors, it is advisable to check and include contributions from other controlling factors as follows:

6. Check contribution from external energy dispersion (Condition 6 with Figure 7). If the situation resembles Figure 7 and the source is located about 25° long upstream, add 20% to the basic value given by 5(c), which will be referred to as the "basic value" hereafter. In case of complications, add 5% to the basic value for each deepening upper vortex within 25° lat. or long. of the incipient disturbance.

7. Check contribution due to stimulation of mass circulation (Figures 4 and 5). Add 5% to the basic value for close resemblance.

8. If vertical wind shear near the centre of the incipient disturbance is greater than 10 kn, reduce the basic value by 10% for each 10 kn shear. (15 taken as 10; 25 and 35 as 30).

It should be borne in mind that Sadler's downward development from the upper vortex is automatically included in the unified model because any upper cyclonic circulation with no corresponding low-level disturbance gives a positive value for $\nabla_h \psi$.

Even if the upper vortex is cold-cored, it can still trigger low-level development unless $\nabla_h \psi - f \nabla_h \phi \leq 0$. However, rapid development must wait until the triggered low-level disturbance brings about sufficient convergence in the moist Ekman boundary layer.

9. The corrected value of D (in units of $10^{-5}$ sec$^{-1}$) may then be rated according to the following criteria for expectant development in the next 24 hours:

1. If D $< 0$, no significant development

2. If 10 $> D > 0$, development to T.D. intensity

3. If 20 $> D > 10$, development to T.S. intensity

4. If 30 $> D > 20$, development to S.T.S. or T. intensity

5. If D $> 40$, rapid development to typhoon intensity

6. If D $> 40$ over a major part of an existing typhoon, watch for development to super-typhoon.

In practice, it is advisable to carry out computations over a network of 9-16 grid points of 2 deg lat. apart around the storm centre and then sketch the isopleth of D to obtain a consistent pattern.
10. For incipient disturbance appearing as a cloud cluster, check whether or not it has become a pre-storm cluster (Figure 23). In the case of a pre-storm cluster with negative or zero D, compute D according to 5(c) as frequently as possible.

11. If contributions from (6) and/or (7) are in favour of development, but $D \leq 0$, keep a close watch for development and compute D according to 5(c) as frequently as possible until $D > 0$.

12. For incipient disturbance in the South China Sea, check with the supplementary procedures. If both systems give comparable results, treat the forecast with confidence. In case of disagreement, downgrade the forecast by one class with the exception of T.D.

Remarks: In using this system, we should be satisfied if the results are often one class out towards lower intensity. For example, if a T.S. is forecast to develop, we may first get a T.D. More failure should be expected in the range $10 \geq D \geq 0$ (all in units of $10^{-5}$ sec$^{-1}$).

* * *
SUPPLEMENTARY PROCEDURES FOR FORECASTING TROPICAL CYCLONE FORMATION IN SOUTH CHINA SEA
(after Brody and Jarrell, 1969)

Worksheet

1. Is there a cyclonic shear zone in the South China Sea on the 12-hourly 850-mb chart?
   Yes  No

2. Is it oriented within 45° of E-W?
   Yes  No
   (If there is a No answer to either #1 or #2, predict no tropical storm formation within 24 to 48 hours)

3. Determine:
   a. Width in ° latitude of belt of east (000° clockwise thru 180°) winds north of cyclonic shear zone
      \[ W_N = \]
   b. Average speed of winds in this northern belt
      \[ V_N = \]
   c. Average direction of winds in this northern belt
      \[ D_N = \]
   d. Width in ° latitude of belt of west (180° clockwise thru 360°) winds south of cyclonic shear zone
      \[ W_S = \]
   e. Average speed of winds in this southern belt
      \[ V_S = \]
   f. Average direction of winds in this southern belt
      \[ D_S = \]

   Determine \( K_N \) by entering table 4 with \( D_N \) and \( W_N \)
   \[ K_N = \]

   Determine \( K_S \) by entering table 5 with \( D_S \) and \( W_S \)
   \[ K_S = \]

   Determine \( K_N V_N \) from table 6 using \( V_N \) and \( K_N \)
   \[ K_N V_N = \]

   Determine \( K_S V_S \) from table 6 using \( V_S \) and \( K_S \)
   \[ K_S V_S = \]

   Add \( K_N V_N \) and \( K_S V_S \)
   \[ S = \]

The parameter \( S \) is a measure of the strength of the cyclonic shear zone.
4. a. Is $S < 20$? 
   Yes No
   b. Is there a tropical storm in the South China Sea? 
      (If either is Yes, predict no new storm formation in South China Sea.) 
      Yes No

5. a. Has $S$ been equal to or greater than 20 on at least two consecutive charts (12-hours apart), including the present one? 
   Yes No
   b. Is the latitudinal variation at 115°E of the cyclonic shear zones of the two latest charts (12-hours apart) less than 4° latitude? 
      If either a or b is No, predict no tropical storm development at this time.
      Yes No
   c. Has $S$ exceeded or been equal to 30 on any chart since it attained a minimum strength of 20 while retaining latitudinal variations less than 4° latitude? 
      If Yes, predict tropical storm development within 24-48 hours within 2° latitude of the cyclonic shear zone. Continue forecasting storm formation until $S < 20$, or until a storm develops in the South China Sea. If no, predict no tropical storm development at this time, but re-examine in 12 hours.

* * *

Acknowledgments

Grateful thanks are due to Mr. G. J. Bell, J.P., Director of the Royal Observatory for valuable advice and enlightening discussions and to Mr. P. C. Chin, Senior Scientific Officer, for constructive criticisms.
References


**TABLE 1**

<table>
<thead>
<tr>
<th>106 km</th>
<th>103 km</th>
<th>10 km</th>
<th>1 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLANETARY</td>
<td>SYNOPTIC</td>
<td>MESO-SCALE</td>
<td>CONVECTIVE OR SMALL SCALE</td>
</tr>
<tr>
<td>LONG WAVES</td>
<td>SUB-TROPICAL DEPRESSIONS</td>
<td>FRONTS</td>
<td>CUMULO-NOBUSTUS SHOWERS</td>
</tr>
<tr>
<td>SUB- TROPICAL ANTI-CYCLOMES</td>
<td>ANTI-CYCLOMES</td>
<td>SQUALL LINES</td>
<td>TORNADOES</td>
</tr>
<tr>
<td>RCI</td>
<td>CLOUD CLUSTERS</td>
<td>MESO-SCALE CONVECTIVE ELEMENTS</td>
<td>CONVECTIVE CELLS</td>
</tr>
<tr>
<td>18 hr</td>
<td>1 hr</td>
<td>1 hr</td>
<td>10 hr</td>
</tr>
</tbody>
</table>

Scales of motion

(after Mason, 1970)
TABLE 2


Units are knots and a positive value denotes cyclonic shear.

| Shear curvature characteristics | Ordinary | Pre-storm | Pre-storm Ordinary | "Clear areas"
|---------------------------------|----------|-----------|--------------------|----------------
| Shear (-∂u/∂y)                  | 10.4     | 9.4       | 9.6                | 3.2            | 17.0 | 2 | -6.8 |
| Curvature (∂v/∂x)               | 1.4      | 2.4       | 2.4                | 1.2            | 1.9  | 5.3 | 3 | -1.6 |
| (Shear/curvature ratio)         | 7.4      | 3.9       | 4.0                | 2.7            | 4.3  | 3.2 | 3/4 | -4.3 |

(adapted from Williams)

TABLE 3

<table>
<thead>
<tr>
<th>Development time</th>
<th>Time taken for central pressure to fall</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1st 20 mb</td>
</tr>
<tr>
<td>Type of tropical cyclone</td>
<td></td>
</tr>
<tr>
<td>SIMULATED (Ooyama's Case A)</td>
<td>84 hr</td>
</tr>
<tr>
<td>NATURAL (Ramage's 7 typhoons of 1965)</td>
<td>5-60 hr (median 31 hr)</td>
</tr>
</tbody>
</table>

(after Ramage, 1970)
TABLE 4

Weighting factor $K_N$ of easterlies determined from the average wind direction $D_N$ of easterly belt and the width $W_N$ of this belt.

<table>
<thead>
<tr>
<th>$D_N$</th>
<th>$W_N$ (° lat.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&gt;6°</td>
</tr>
<tr>
<td>000° - 124°</td>
<td>1.00</td>
</tr>
<tr>
<td>125° - 180°</td>
<td>0.50</td>
</tr>
</tbody>
</table>

TABLE 5

Weighting factor $K_S$ of westerlies determined from the average wind direction $D_S$ of westerly belt and the width $W_S$ of this belt.

<table>
<thead>
<tr>
<th>$D_S$</th>
<th>$W_S$ (° lat.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&gt;6°</td>
</tr>
<tr>
<td>180° - 304°</td>
<td>1.00</td>
</tr>
<tr>
<td>305° - 360°</td>
<td>0.50</td>
</tr>
</tbody>
</table>

TABLE 6

Partial strengths $K_N V_N$, $K_S V_S$, of cyclonic shear zone derived from average wind speeds $V_N$, $V_S$ and weighting factors $K_N$, $K_S$. Strength $S$ is the sum of the partial strengths $V_{N,S}$.

<table>
<thead>
<tr>
<th>$K_{N,S}$</th>
<th>0</th>
<th>5</th>
<th>10</th>
<th>15</th>
<th>20</th>
<th>25</th>
<th>30</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.15</td>
<td>0</td>
<td>0.8</td>
<td>1.5</td>
<td>2.2</td>
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</table>
Figure 1 - Schematic representation of trade-wind - ITCZ circulation showing the structure of the associated convective cloud systems in vertical cross-section (after Mason, 1970)

Figure 2 - Illustration of cluster classification scheme (after Williams, 1970)
Figure 3 - Vertical shear of zonal wind \( u_{950} - u_{200} \) at 4° latitude intervals across cluster centres. Westerly shear denotes stronger west winds (or weaker east winds) aloft than near surface

(after Williams, 1970)
Figure 4 - Role of the upper atlantic tropospheric shear line in stimulation and/or the constraint of mass circulation in a tropical cyclone. As the cyclone approaches the wind maximum south of the shear line, mass circulation in the cyclone is stimulated. As the cyclone passes between the wind maximum to the south and that to the north of the shear line, mass circulation is constrained

(after Simpson, 1970)
Figure 5 - Influence of the upper troposphere shear line on the development of hurricane "Holly". As "Holly" initially approached the shear line, it rapidly intensified to hurricane force. In this instance the shear line provided a field of anticyclonic shear as an environment for the outflow of the cyclone. Subsequently, as the cyclone proceeded westward to a point where the shear line provided an environment of cyclonic shear for the outflow, the cyclone rapidly lost intensity.

(after Simpson, 1970)
Figure 6 (a) - Portrayal of how cyclonic shear in a zonal non-divergent trade wind current at 950 mb can produce sub-cloud convergence if a frictional veering of $10^\circ$ were present. $V_{sfc}$ is the meridional surface wind (after Gray, 1968)

Figure 6 (b) - Idealized portrayal of the difference in wind directions at the surface and at the top of the friction layer relative to a doldrum Equatorial Trough. Note that the wind south of the Equatorial Trough is in general weak and that the sharp cyclonic gradient of trade wind on the poleward side of the Equatorial Trough can lead to substantial low-level convergence by virtue of frictional veering (after Gray, 1968)
Figure 7 - Schematic streamline representation of a common west Pacific sequence at 30,000 or 40,000 ft leading to typhoon development. Cyclonic vorticity increases in Trough AA, resulting in downstream energy dispersion which sharpens ridge BB and intensifies the next downstream trough CC. The pressure fall along CC may trigger any low-level cyclonic disturbance in the shaded area into a typhoon

(after Ramage, 1959)
(a) Characteristic Scale of Adjustment Towards Thermal Wind Balance (L<sub>n</sub>) as a function of latitude and environmental lapse rate (γ) in excess of saturated adiabatic lapse rate (γ<sub>m</sub>).

\[ L_n = \frac{RT}{\sqrt{T_n - T_m}} \]

γ = 260°F (-13°C)

L<sub>n</sub> = ∞ at Equator

(b) Tephigram portrayal of the mean summertime tropical radiosonde in regions where tropical storms develop. Numbers in the left center portion of figure represent parcel ascent beginning at the pressure levels indicated. Note that for parcel ascent from levels higher than 900 mb, the air parcel remains cooler than its environment.

(after Gray, 1968)

(c) Same as for (b) but for average conditions within the rain area of the typical tropical disturbance which later becomes a tropical storm. Only parcel ascent from levels below 850 mb will produce warming.

(after Gray, 1968)

Figure 8
FIGURE 9. ASSUMED AXIAL SYMMETRY IN 250 mb - 750 mb WIND SHEAR IN THE INITIAL DISTURBANCE
INITIAL 250 - 750 mb MEAN TEMPERATURE (THICKNESS)
FIELD SATISFIES $\nabla^2_h (\Psi_{250} - \Psi_{750}) > f^{-1} \nabla^2_h (\Phi_{250} - \Phi_{750})$
Figure 10. Computed development of a weak incipient disturbance into an intense vortex in about 24 hr.

(After Chen, 1964)
Table 1: Specifications of Case A.

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<th>Symbol</th>
<th>Section where defined</th>
<th>Assumed value</th>
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<td>$r_e$</td>
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<td>$\beta$</td>
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<td>$C_D$ variable: Eq. (9.1)</td>
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<td>$\eta$</td>
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Figure 11 - Schematic diagram depicting the basic design of the model.

Figure 12 - Development of computed tropical cyclone in Case A. Upper left: time, radial, and central wind profiles; mid-left: wind speed at radius $r$; right: central pressure vs. time.

Figure 13 - Radial distribution of various field variables from 0 to 400 km in Case A at $t = 134$ hr.

(after Ooyama, 1969a)
Figure 14 - Variation with time of: (a) max \( v_i \), (b) the radius of max \( v_i \) and inner and outer limits of hurricane force winds, and (c) the minimum surface pressure (deviation from the normal). Solid lines are computed with Model II, and dashed lines with Model I.

Figure 15 - Comparison of Model II cyclones computed with different \( \Delta r \) and different difference schemes.

Figure 16 - Radial distribution at \( t = 146 \) hr of: (a) \( v_0 \) and \( v_i \), and (b) \( w \). Solid lines are computed with Model II, and dashed lines with Model I.

(after Ooyama, 1969b)
(a) Effects of the sea surface temperature on the intensity and size of computed cyclones. In Cases A and B, $T_s$ is held constant at 27.5 and 25.6°C, respectively. In Case AB, $T_s$ is suddenly decreased from 27.5 to 25.6°C at $t=134$ hr with an opposite change in Case BA at $t=104$ hr. The time origin of integration in Case B is at $-100$ hr (off the diagram).

(b) Results of experiments demonstrating the importance of direct energy supply from the oceans to tropical cyclones. In Case AE, the energy supply to the Case A cyclone is suddenly stopped ($C_p=0$) at $t=134$ hr, to simulate landfall. In Cases $E_6$ and $E_9$, in which $v$ is initially assumed to be 2.0 and 3.0, respectively, there is no energy supply from the ocean.

(c) Effects of sea surface temperature distribution on cyclone development. In Cases $C_6$, $C_7$, and $C_9$, $T_s=25.6$°C for $r>300$ km, 23.5°C for $r>150$ km, and 23.9°C for $r>150$ km, respectively, while $T_s=23.5$°C in the inner region in all cases. In Case A, the warm area extends to 1000 km.

Figure 17
(after Ooyama, 1969b)
Figure 18 - Development of computed tropical cyclones from weak initial vortices of various sizes. Initially, \( \text{max } v_i \) is 5 m sec\(^{-1}\) in all cases and radii of \( \text{max } v_i \) are 25, 50, 75 and 100 km in Cases \( Ai_1 \), \( Ai_2 \), \( Ai_3 \), and \( Ai_4 \), respectively. The time origin of integration in the last three cases is shifted as indicated for easier comparison. The lateral eddy viscosity is very small in Case \( Ai_1 \) which is otherwise identical to Case \( Ai_1 \) in specification (after Ooyama, 1969 a).
Figure 19 - Rate of central surface pressure falls in
1) Ooyama's (1969) model tropical cyclone
(from his Figure 4), the heavy full line;
and 2) seven disturbances which developed
into intense typhoons over the western
Pacific during 1965 (Fleet Weather Central/
Joint Typhoon Warning Center, 1965). When
central pressure falls to about 1000 mb in
this region, winds reach about 10 m sec⁻¹
(Ooyama's initial condition)

(after Ramage, 1970)

Figure 20 - Top: Radii of the
maximum surface wind
and maximum wK,
Outer and inner
limits of gale
and hurricane force
winds. Middle: Central
pressure as a function
of time. Bottom:
Maximum surface wind
as a function of time

Figure 21 - Radial distributions of
variables at selected
times. Left: Surface
wind speed. Centre:
Surface pressure.
Right: Vertical motion
at the 1054 meter level
Figure 22 - Vertical cross-sections of various dependent variables at 180 hours. **Upper left**: Tangential velocity in meters/sec. **Upper right**: Total wind speed in meters/sec. **Centre left**: Radial velocity in meters/sec. **Centre right**: Temperature excess over mean tropical atmosphere in degrees Kelvin. **Lower left**: Vertical velocity in meters/sec. (Note isopleths are drawn at variable intervals.) **Lower right**: Pressure departures from mean tropical atmosphere in millibars. (Note isopleths are drawn at variable intervals)

(after Rosenthal, 1969)
Figure 23—Various Phases of Typhoon (Hurricane) Development in Relation to Pre-requisite Conditions and Triggering Mechanisms

MPT: Mid-Pacific Trough; TTL: Tropical Upper Tropospheric Trough
APPENDIX A

VORTICITY SCALE FROM CONTOURS

\[ \zeta_0 = \left[ \Phi_1 + \Phi_2 + \Phi_3 + \Phi_4 - 4 \Phi_0 \right] \frac{gT}{\lambda^2} \quad \text{for contours in geopotential metres} \]

UNIT: \(10^{-5} \ \text{sec}^{-1}\) FOR CONTOURS IN GEOPOTENTIAL METRES

VALID FOR 1:10 m MERCATOR PROJECTION

\[ \frac{\zeta v}{\zeta n} \] VORTICITY SCALE FROM STREAMLINES & ISOTACHS

UNIT: \(10^{-6} \ \text{sec}^{-1}\) FOR SPEED IN KNOTS

VALID FOR 1:10 m MERCATOR PROJECTION
APPENDIX B

\( \zeta = \delta v / \delta n + cV \)

UNIT: \( 10^{-6} \text{ sec}^{-1} \)

FOR SPEED IN KNOTS

VALID FOR 1:10m MERCATOR PROJECTION