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THE GROWTH OF TROPICAL CYCLONES IN THE SOUTHWEST PACIFIC

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Abstract

The concept of intensification potential previously presented in climatological terms, is applied to individual disturbances using a new definition of tropospheric vertical wind shear. Also discussed is the effect on development of the meridional component of motion.

A prediction diagram is constructed which may prove a useful forecasting tool for determining the likelihood of a circulation ultimately becoming a severe storm and for estimating rate of development.

Introduction

Until recently there has been lack of agreement on the environmental conditions and physical processes necessary for the development of tropical cyclones. Charney and Eliassen (1964) proposed conditional instability of the second kind (CISK) as a mechanism by which frictionally-forced convergence in the boundary layer in conjunction with the heating potential of cumulus convection initiates the development of tropical cyclones.

Mendenhall (1967) and Gray (1968) have shown in statistical studies that significant Ekman or frictionally induced wind veering does in fact occur in the sub-cloud layer (lowest 600-700 metres) over the tropical oceans. In this way the cyclonic wind shears will produce convergence in the planetary boundary layer and vertical motion at the top of this layer. This mechanism transports water vapour into the troposphere where warming takes place by the release of the latent heat of condensation in the deep cumulus and cumulonimbus.

Yanai (1961) considered that the condensation process does not actually contribute to the production of the primary vertical circulation in tropical cyclones even at the formative stage but that its importance lies rather in the creation and maintenance of the warm core. This heating should be regarded only as potential heating for unless it is concentrated in a small area it will not provide the necessary tropospheric warming for the production of the warm core and the resulting pressure fall. Gray (1970) and Simpson (1970) both observe that lack of ventilation, shown by small or zero vertical wind shear, accompanies cyclone formation.

Williams (1970) has made an extensive statistical study of satellite cloud clusters in the western tropical North Pacific. All clusters were found to be associated with small values of vertical tropospheric wind shear. Clusters which subsequently developed into storms exhibited large low-level cyclonic shears, whereas the remaining cloud categories showed weaker cyclonic shears.

The most recent summary of observed storm development is due to Gray (1968). Following Hutchings (1953) and Gabites (1956, 1963) he notes that most tropical cyclones in the Southwest Pacific originate in the region depicted in Fig. 1. This region produces an average of 7 tropical cyclones per year, or 11% of the global total, and in agreement with Palmén (1948, 1956) lies well within the 26.5°C sea surface isotherm for January.

The importance of sea surface temperature lies in its strong influence on the potential buoyancy of cumulus, defined by Gray (1968) as the difference between the equivalent potential temperature at the surface and at 500 mb. Figure 1 shows that the largest average values of potential buoyancy in January and February, correspond closely with the region of main storm development. Potential buoyancy plays an important role in determining the degree to which cumulus and cumulonimbus can warm the middle and upper troposphere by condensation heating. The ventilation of a cluster of cumulus and cumulonimbus is strongly correlated with the vertical shear and the degree of condensation heating is dependent on the surface convergence, or surface relative vorticity.

If frictionally induced surface convergence (and potential heating) is directly related to surface relative vorticity as is required from Ekman theory, then the ratio of surface relative vorticity (ζ_r) to tropospheric vertical wind shear $|S_z|$ would be a representative parameter of actual heating, (Gray 1968). Climatologically, places having the maximum value of the ratio $\frac{\zeta_r}{|S_z|}$ are correlated with the locations of tropical disturbance genesis and initial disturbance intensification (Fig. 2).

The data network in the Southwest Pacific allows reasonable daily estimates to be made of sea level vorticity but not of potential buoyancy. The value of the ratio $\frac{\overline{\zeta_r}}{\overline{|S_z|}}$, defined as the Intensification Potential, I, has been calculated for a number of incipient disturbances, and in this paper an attempt is made to relate it to the subsequent development of individual storms.

The evaluation of I

(a) The tropospheric mean vertical wind shear

The tropospheric mean vertical wind shear may be calculated directly using streamline-isotach analyses for the 1000, 850, 700, 500, 300 and 200 mb levels. Following Ward (1971a) the mean zonal flow $\bar{u}$, throughout a column extending from 1000 mb to 100 mb and having an area of 5 degrees latitude centred on the disturbance is given by

$$\bar{u} = \frac{1}{36}(3u_{10} + 6u_{85} + 7u_7 + 8u_5 + 4u_3 + 8u_2)$$

where u_{10}, u_{85} etc. represent the mean flow at 1000, 850 mb etc.

If the lower troposphere extends from 1000 mb to 550 mb and the upper from 550 mb to 100 mb, it is easily shown that

$$\bar{u}_B \doteqdot \frac{1}{18}(3u_{10} + 6u_{85} + 7u_7 + 2u_5)$$

and $$\bar{u}_T \doteqdot \frac{1}{18}(6u_5 + 4u_3 + 8u_2)$$

where $\bar{u}_B$ and $\bar{u}_T$ are the mean values of the zonal flow in the lower and upper halves of the troposphere respectively.

The tropospheric mean vertical shear of the zonal wind is then given by

$$\bar{u}_z = \bar{u}_T - \bar{u}_B$$

i.e. $$\bar{u}_z \doteqdot \frac{1}{18}(8u_2 + 4u_3 + 4u_5 - 7u_7 - 6u_{85} - 3u_{10})$$

A similar expression is obtained for the tropospheric mean vertical shear of the meridional wind, $\bar{v}_z$.

(b) Mean relative vorticity of disturbances

For disturbances that have not reached the cyclone stage mean values of surface relative vorticity may be calculated using mean values of u and v obtained by averaging over the same 5 degree latitude grid used in obtaining the tropospheric mean vertical wind shear.

If it is assumed that in disturbances which have achieved some degree of cyclonic symmetry,

$Vr^{\frac{1}{2}}$ = constant within 2° latitude of the storm centre, (Byers, 1944; Hughes, 1952) then,

$\overline{\zeta}_r = \frac{1}{2}\frac{V}{r}$, where V is wind speed at distance r from the storm centre.

If it is further assumed that the maximum sustained wind around the storm occurs at 30 nm distance from the centre

$$\overline{\zeta}_r = \frac{V}{2.16} \times 10^{-5}\ sec^{-1}$$, where V is expressed in knots.

Values of $\overline{\zeta}_r$ for various wind speeds are given below

V	20	30	40	50	60	70	80	90	100	Knots
$\overline{\zeta}_r$	9	14	19	23	28	32	37	42	46	$X10^{-5} sec^{-1}$

Use of the Intensification Potential, I, in predicting the development of disturbances

Gray (1968) summarises the primary requirements for tropical storm development as:

(i) climatological requirement - sea surface temperature above 26-27°C etc.;

(ii) general circulation requirement - existence of synoptic-scale cyclonic horizontal wind shear;

(iii) area size and low level convergence requirement - large horizontal sea level wind shear and convergence;

(iv) time requirement - low level convergence must act over a sufficiently long period;

(v) ventilation requirement - small vertical wind shear;

(vi) vertical momentum transfer requirement - cumulus up and down drafts must inhibit increase of vertical shear as low level convergence increases with increasing curvature and wind speed.

Provided that an incipient storm lies within a region satisfying the climatological requirement, the requirements (ii) to (v) will be satisfied, and further development may be expected to follow, if the ratio $I = \frac{\overline{\zeta}_r}{|S_z|}$ is relatively high over the disturbance area. Requirement (vi) should be satisfied if the trend in values of I, at successive observation times, is not towards lower values.

The mean values of the Intensification Potential I, during the developing stage of 18 tropical disturbances in the Southwest Pacific, are listed in Table 1. All disturbances developed north of 20°S, and their tracks are shown in Figs 3,4. The track of disturbance "Dolly" was unusual (Hill, 1970; Ward, 1971b) and in this study the three distinct phases of its movement, firstly towards the eastsoutheast, secondly towards the northnortheast and thirdly towards the southeast, have been treated separately.

Nearly all of the disturbances that developed and eventually reached storm or hurricane intensity were characterised by a relatively high mean value of I during their developing stage. In all except one case the mean value of I exceeded 3.0.

Disturbances that failed to develop were characterised by low mean values of I, all less than 2.0.

<u>The observed influence of the meridional component of movement, $\bar{v}$, of the disturbance on development</u>

In attempting to relate the future 24-hour vorticity change, or change in the maximum sustained wind of the disturbance to the intensification potential I, it was observed that the meridional component of motion, $\bar{v}$ appeared to be related to the development of the disturbance. Poleward moving disturbances generally intensify in the tropics while those with an equatorward component to their motion tend to weaken. However equatorward motion is comparatively rare and may at times be accompanied by intensification. Instantaneous values, at 0000Z, of I and $\bar{v}$ for the 18 disturbances, 53 days in all, are shown in Fig. 5. Most of those combinations of I and $\bar{v}$ accompanying storm development fall into one group, and those combinations with which there is no significant development fall into another. There are only a few combinations common to both groups (Fig. 5).

The number alongside each plot in Fig. 5 gives the average rate, in knots per day, at which the maximum sustained wind in the disturbance increased. These figures are only approximate since direct observations of maximum winds were not always available. The averaging process also implies a linear increase of wind speed with time.

Variations in the value of I in the "storm likely" area of Fig. 5 do not appear to affect the rate of development but the latter is greater than 5 knots/day throughout this area and averages 10 knots/day. It appears rather that a critical value of I exists (about 2.4 for $\bar{v}$ = 0) above which storm development is virtually certain. The influence of the meridional component of motion, $\bar{v}$, on development, appears to be comparatively small in the tropics. (See Appendix A).

It is significant to mention that the March 1968 cyclone (mean I = 2.4) formed in a latitude somewhat south of the normal area of initial formation, and it is conjectured that baroclinic processes may have had considerable influence on its development.

Furthermore, sea-surface temperatures and potential buoyancy are not explicitly taken into account in the calculations of I. The assumption that these remain at fixed values for all disturbances in the development area may lead to considerable variations in the "effectiveness" of I for individual disturbances.

Conclusions

The results obtained in the present work, although taken from a very small sample, indicate that the 'Intensification Potential' concept combining the effect of surface relative vorticity and vertical shear is a useful predictor of cyclone development when applied to individual disturbances. A critical value of Intensification Potential appears to exist (about 2.4) above which cyclone development is almost certain and below which little development is likely. However, due to neglect of sea-surface temperature and potential buoyancy in the calculations this value probably varies slightly from disturbance to disturbance.

Furthermore, it appears that the influence of the meridional component of motion, although small in the tropics, increases the effectiveness of Intensification Potential for poleward motion and decreases it for motion towards the equator.

In conjunction with the satellite information currently more readily available, it is now possible by the above technique to distinguish at an early stage cloud clusters which are likely to grow into tropical storms from those which should not develop.

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TABLE 1.

Mean values of Intensification Potential, I, and Meridional component of motion, $\bar{v}$, (measured during the main developing stage, i.e. north of 20°S.) and corresponding mean values of 24-hour change in strength of maximum sustained wind, $\Delta \bar{V}$, (measured from time of initial detection of disturbance to time of eventual maximum development).

Disturbance (Date of initial identification)		Mean I (c.g.s. unit 10^{-7})	Mean $\bar{v}$ (knots)	Mean $\Delta \bar{V}$ (knots)	Eventual maximum development
12.2.70	"Dolly" (phase 2)	46.0	+ 5	+ 8	Hurricane (70 kts)
28.3.67	"Glenda"	13.7	- 8	+15	Storm (60 kts)
12.2.70	"Dolly" (phase 3)	9.3	- 3	+11	Hurricane (70 kts)
7.4.67		7.7	- 6	+ 7	Hurricane (65 kts)
28.2.70	"Emma"	7.5	- 4	+10	Hurricane (80 kts)
23.1.67	"Dinah"	7.5	- 5	+ 6	Storm (60 kts)
7.4.70	"Gillian"	5.8	- 4	+ 8	Hurricane (64 kts)
12.2.70	"Dolly" (phase 1)	5.0	- 5	+ 8	Storm (60 kts)
6.3.63		4.8	+ 4	+ 9	Hurricane (65 kts)
12.1.69		4.2	- 8	+ 7	Storm (50 kts)
5.4.68		4.0	-12	+14	Hurricane (100 kts)
30.1.69	"Colleen"	3.1	-10	+11	Hurricane (80 kts)
2.3.68		2.4	- 2	+20	Hurricane (100 kts)
13.4.70	"Helen"	2.2	- 4	+ 4	Gale (40 kts)
4.1.70	"Alice"	1.9	- 2	+ 1	Small disturbance (20 kts)
8.1.70	"Bonnie"	1.0	- 8	+ 2	Small disturbance (25 kts)
9.1.70	"Claire"	1.0	- 2	+ 1	Small disturbance (20 kts)
9.2.70		0.7	- 4	+ 1	Small disturbance (25 kts)

APPENDIX

THE EFFECT OF THE MERIDIONAL COMPONENT OF MOVEMENT, $\bar{v}$, OF THE DISTURBANCE, ON DEVELOPMENT

For normal synoptic scale motions in the atmosphere (including the tropics) the simplified vorticity equation may be written

$$\frac{d\zeta}{dt} = -\beta v - (f + \zeta)\,\nabla\cdot\underset{\sim}{V} \qquad \text{------------} \quad (1)$$

where $\frac{d\zeta}{dt}$ = rate of change of vertical component of relative vorticity, ζ, following the horizontal motion,

$\beta = \frac{\partial f}{\partial y}$, variation of Coriolis parameter, f, with latitude

$\underset{\sim}{V}$ = the horizontal wind,

$\nabla\cdot\underset{\sim}{V}$ = horizontal divergence $= \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} - \frac{v}{a}\tan\phi$

ϕ = latitude

and a is radius of the earth.

Treating the cyclone vortex as a "small parcel" of air and assuming that the meridional component of motion of the vortex represents the corresponding motion of the "small parcel", which follows from tropospheric mean flow theory (Ward 1971), a necessary condition for the intensification of a cyclone (ζ positive) in the northern hemisphere is that

$$\beta v + (f + \zeta)\,\nabla\cdot\underset{\sim}{V} < 0$$

We now consider the following special cases of cyclone motion:

(i) Poleward motion, flow convergent

It is easily verified that the two terms on the R.H.S. of (1) oppose one another, the $(f+\zeta)\,\nabla\cdot\underset{\sim}{V}$ term tending to increase cyclonic vorticity while the βv term tends to decrease cyclonic vorticity. Cyclonic development will be enhanced if

$$\left|(f + \zeta)\,\nabla\cdot\underset{\sim}{V}\right| > \left|\beta v\right|$$

and inhibited if

$$\left|(f + \zeta)\,\nabla\cdot\underset{\sim}{V}\right| < \left|\beta v\right|$$

(ii) Equatorward motion, flow convergent

The two terms on the R.H.S. of (1) now support one another in tending to increase cyclonic vorticity.

(iii) Divergent flow tends to inhibit cyclonic development in both poleward and equatorward motion.

The meridional component of motion, v , of a parcel of air contributes an amount $-\frac{v}{a}\tan\phi$ due to the convergence of the meridians, to the total divergence. Near the equator $\tan\phi \doteqdot 0$ but increases rapidly as the parcel moves poleward, so that the contribution of $-\frac{v}{a}\tan\phi\,(f+\zeta_r)$ to the vorticity change becomes increasingly significant.

If at latitude 15°, where $f = 3.775 \times 10^{-5}\ sec^{-1}$, $v = 5$ mps, for a disturbance with maximum sustained wind of 30 kts $\frac{v}{a}\tan\phi\,(f+\zeta_r) \doteqdot 3.7 \times 10^{-11}\ sec^{-2}$.

For a disturbance at the same latitude with maximum sustained winds of 90 kts the value of

$$\frac{v}{a}\tan\phi\,(f+\zeta_r) \doteqdot 9.6 \times 10^{-11}\ sec^{-2}.$$

At latitude 15°, βv which tends to decrease cyclonic vorticity in poleward flow is approximately $11 \times 10^{-11}\ sec^{-2}$.

The net contribution to vorticity change by the two terms involving v is, for poleward moving disturbances at latitude 15°

$-7.3 \times 10^{-11}\ sec^{-2}$ for a small disturbance

and $-1.3 \times 10^{-11}\ sec^{-2}$ for a large disturbance,

while at latitude 30° the net contribution to cyclonic vorticity production is almost zero for a small disturbance and increases to $+12.4 \times 10^{-11}\ sec^{-2}$ for a large disturbance.

Since the total 24 hour vorticity change in a developing tropical storm is $1 \sim 4 \times 10^{-9}\ sec^{-2}$ the terms involving v do not contribute more than 10%. This is in agreement with the empirical results presented in Fig. 5.

The complete vorticity equation includes "tilting" or vortex tube and friction terms. The value of the term arising from the turning of vortex tubes is of the same order of magnitude as $\frac{v}{a}\tan\phi\,(f+\zeta_r)$. The value of the friction term is unknown but it is probably significant especially in vigorous storms.

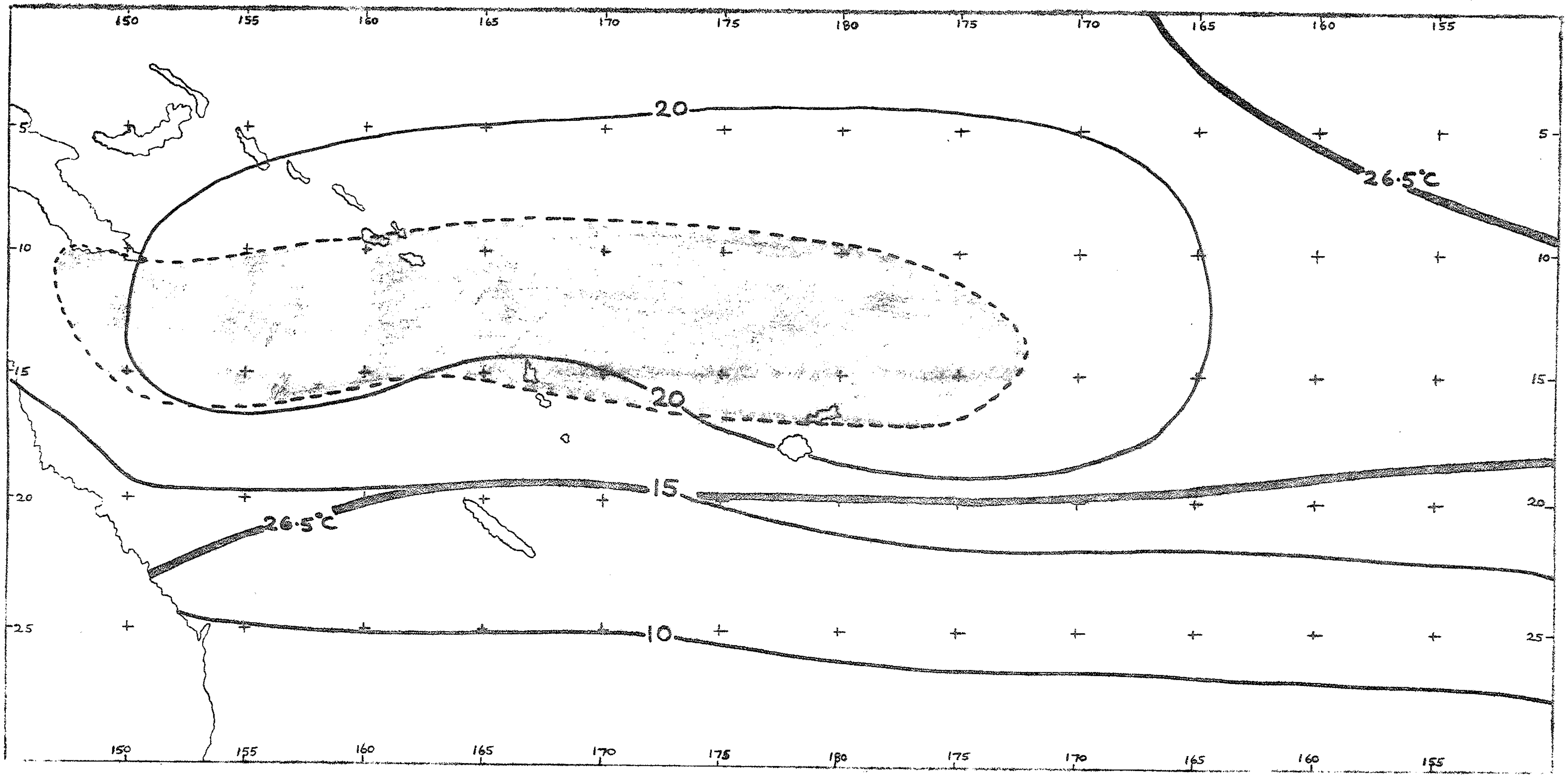

Fig. 1 Mean potential buoyancy of cumulus for January - February. Shaded area depicts main storm development region. Thick line is 26.5°C isotherm for January. (Mainly after Gray).

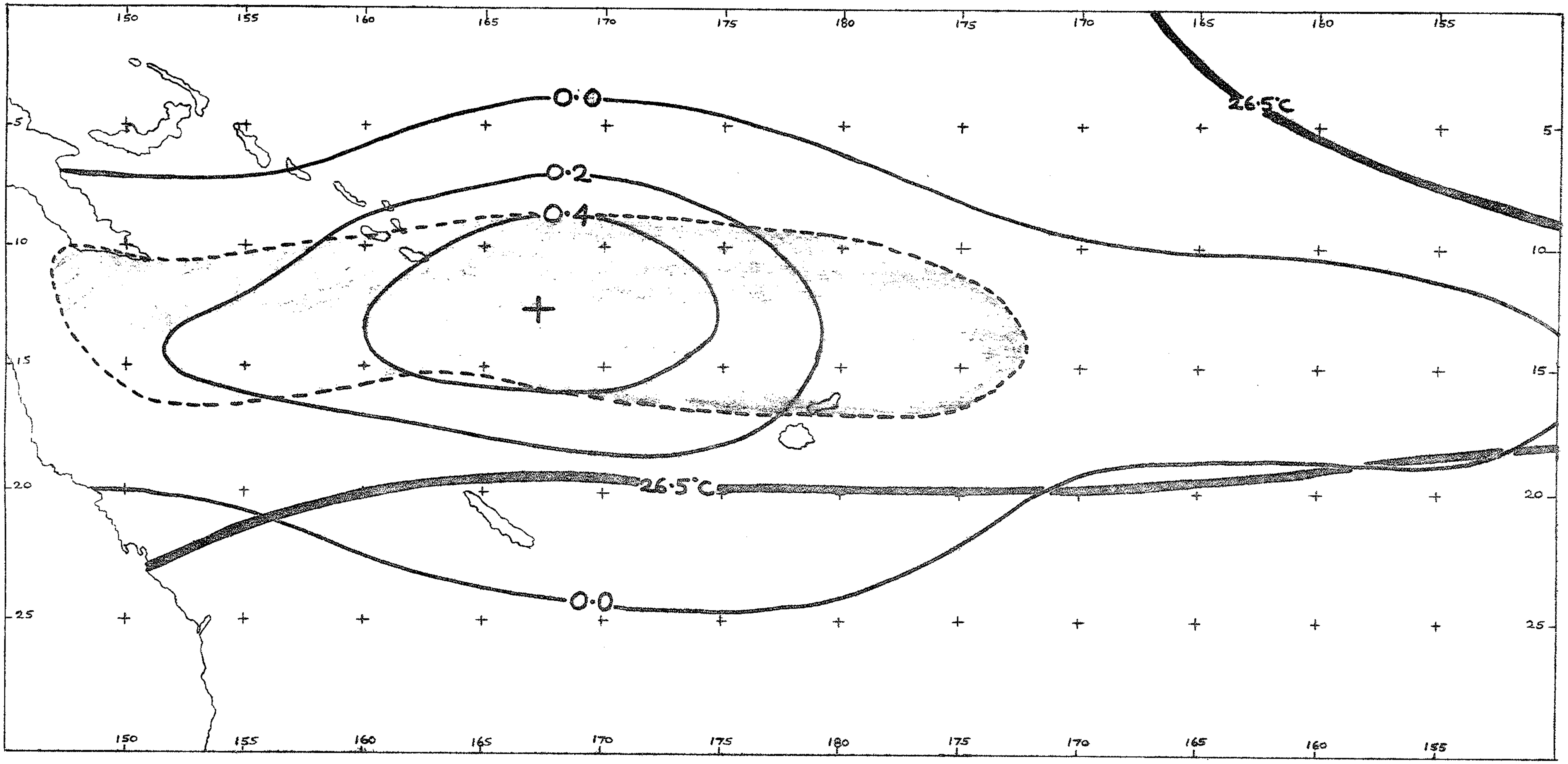

Fig. 2 Mean summer month distribution of Intensification Potential, I (c.g.s unit 10^{-7}). Shaded area depicts main storm development region. Thick line is 26.5°C isotherm for January. (Mainly after Gray).

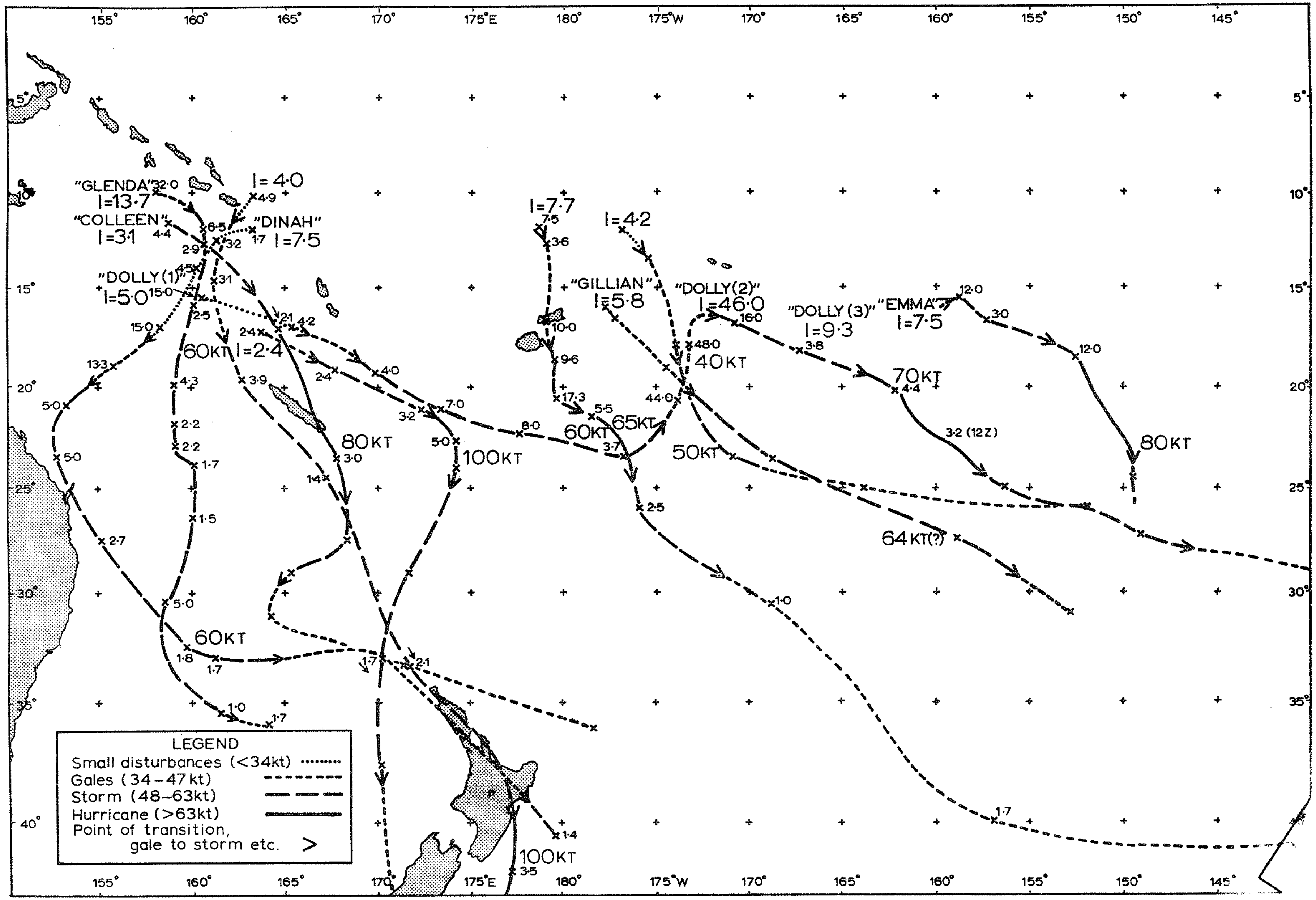

Fig. 3 Tracks of disturbances that developed to storm or hurricane intensity. 0000 GMT positions shown by X. Mean values of I during developing stage are shown near the commencement of each track and daily values are shown alongside 0000 GMT positions.

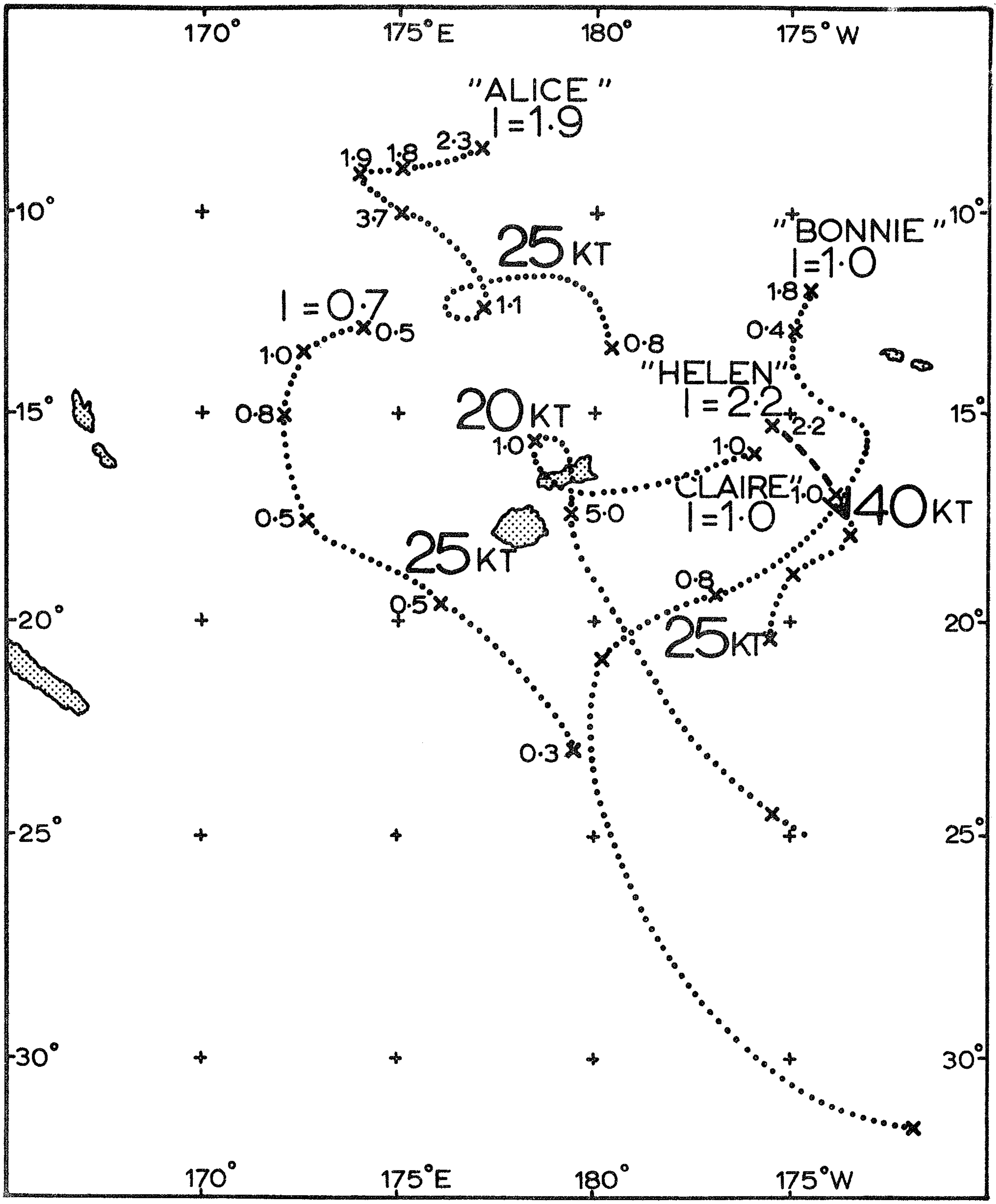

Fig. 4 Tracks of disturbances that failed to develop beyond gale intensity. (For legend see Fig. 3)

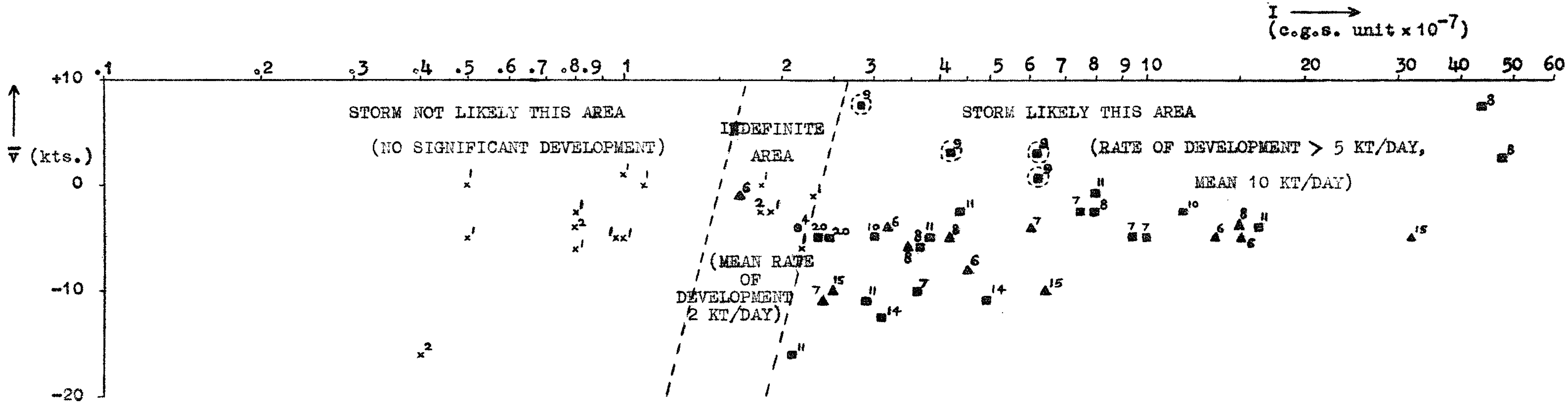

Fig. 5 Plot of instantaneous values of Intensification Potential, I, versus meridional component of motion, $\bar{v}$, for individual disturbances while in the main development area (i.e. equatorward of 20°S.). Ultimate maximum development and mean rate of intensification in knots/day (see text) denoted by:

$\times^1$ small disturbance

$\bullet^4$ gale

$\blacktriangle^{10}$ storm

$\blacksquare^{10}$ hurricane