On the Meteorological Elements' Distribution in the Region of the Typhoon (I)

By S. Tomitaka

Fukuoka Meteorological Observatory

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Abstract

A formula of gradient wind velocity and relative vorticity for typhoon is derived, assuming that the pressure distribution at sea level is circular in the region of typhoon.

And the distribution of the wind velocity is derived by introducing the values from the above-mentioned formula to the general solution of equation of motion in the frictional layer as the function of the height and the distance from the center.

The height of frictional influence and the value of the other elements are estimated for the region of the typhoon.

These calculating methods are tested with the synoptic data about the Typhoon No. 12 (June) which invaded Kyushu district in the year 1954.

1. Introduction

The lives of several thousand peoples and the materials of hundred milliard yen were deprived by an inversion of violent Typhoon Isewan at Nagoya district in a night last year.

We can not but be astonished how mighty is Nature's violence now.

From a view point of preventing the disasters, the study of typhoon became important increasingly in the recent years.

In the first stage, it was pointed out by Dr. Okada, Dr. Fujiwara Dr. Horiguchi and Dr. Namekawa being responded by many students and the excellent researchs have been performed in the late years in our country. While, the Meteorological Agency installed an electronic calculating machine last year.

And so, the numerical prediction of the typhoon course is expected to develop largely and rapidly.

In this reports, firstly, it is endeavoured to get the equation of pressures' distribution correctly, as Dr. Horiguchi, Dr. Takahashi, Dr. Fujita and Mr. Masuda derived already.

Secondary, the author got the equation of gradients' wind velocity and relative vorticity using the equations of pressures' distribution obtained firstly.

The general solution of the equation about the motion of atmosphere with frictional layer has already been derived by Dr. Syono. Here, the equation is expressed by a cylindrical coordinate, being assumed stationary, and is balanced with Coriolis force, pressures' gradient and frictional force.

Then, the distribution of wind velocity in the region of the typhoon can be obtained as the function of height z and the distance from the center r , by introducing above-cited relative vorticity and gradient wind velocity to these general solution.

And using the values thus obtained, the distribution of horizontal divergence, ascensional rate and eddy precipitation can be obtained as the function of height z and the distance from the center r too.

It is very convenient for these methods that the observed data of sea level pressure is more numerous than the data of upper layers, and that aircraft observation is enforced for every typhoons about the centers' position and pressure, wind velocity, etc.

2. Theoretical considerations for the distributions of each meteorological element in the region of typhoon

A distribution of pressure in the region of typhoon is assumed generally shown by,

$$p(r) = p(\infty) - \frac{4p}{n} \sum_{j=1}^n a_j \exp\left(-\frac{r^m}{r_j^m}\right), \quad (1)$$

where, $p(r)$: pressure in the distance of r ,
 $p(\infty)$: pressure of undisturbed environment,

$4p$: the depth of the pressure in the center,

a_j, r_j, m, n : constants which decide the form of pressure distribution in the region of typhoon.

Differentiating the equation (1) in respect of r , we have,

$$\frac{\partial p(r)}{\partial r} = \frac{4p}{n} \sum_{j=1}^n a_j m \frac{r^{m-1}}{r_j^m} \exp\left(-\frac{r^m}{r_j^m}\right), \quad (2)$$

generally, the gradient wind velocity v_g is given as,

$$v_g = -\frac{1}{2}fr + \left\{ \left(\frac{1}{2}fr \right)^2 + \frac{r}{\rho} \left(\frac{\partial p}{\partial r} \right) \right\}^{\frac{1}{2}}, \quad (3)$$

where, $f = 2\omega \sin \varphi$: Coriolis parameter,

$\omega = 7.29 \cdot 10^{-5}$ rad sec $^{-1}$: angular velocity of rotation of the earth,

ρ : density of air.

Substituting (2) into (3), we have,

$$v_g = -\frac{1}{2}fr + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho n} \sum_{j=1}^n a_j m \left(\frac{r}{r_j} \right)^m \times \exp\left(-\frac{r^m}{r_j^m}\right) \right]^{\frac{1}{2}}. \quad (4)$$

Neglecting the higher order, the relative vorticity ζ is,

$$\zeta = \frac{\partial V}{\partial r} + \frac{V}{r}. \quad (5)$$

Consequently, the relative gradient vorticity ζ_g would be,

$$\begin{aligned} \zeta_g = & -f + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho n} \sum_{j=1}^n a_j m \left(\frac{r}{r_j} \right)^m \right. \\ & \times \exp\left(-\frac{r^m}{r_j^m}\right) \left. \right]^{-\frac{1}{2}} \\ & \times \left[\frac{1}{2}f^2r + \frac{4p}{\rho n} \sum_{j=1}^n a_j m \frac{r^{m-1}}{r_j^m} \right. \\ & \times \left. \left\{ 1 + \frac{m}{2} - \frac{mr^m}{2r_j^m} \right\} \exp\left(-\frac{r^m}{r_j^m}\right) \right]. \quad (6) \end{aligned}$$

Then, a simplified case of above-stated equations shall be considered;

(i) if we put these constants as $n=1$, $a_1=1$, $m=1$, $r_1=N$, then equations (1), (4) and (6) would be deduced to,

$$\frac{4p(r)}{4p} = \exp\left(-\frac{r}{N}\right), \quad (7)$$

$$v_{gI} = -\frac{1}{2}fr + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho} \frac{r}{N} \exp\left(-\frac{r}{N}\right) \right]^{\frac{1}{2}}, \quad (8)$$

and,

$$\begin{aligned} \zeta_{gI} = & -f + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho} \frac{r}{N} \exp\left(-\frac{r}{N}\right) \right]^{-\frac{1}{2}} \\ & \times \left[\frac{1}{2}f^2r + \frac{4p}{2\rho N} \left(3 - \frac{r}{N} \right) \exp\left(-\frac{r}{N}\right) \right], \quad (9) \end{aligned}$$

(ii) if we put these constants as $n=2$, $a_1=a_2=1$, $r_1=L$, $r_2=M$, $m=2$, then equations (1), (4) and (6) would be deduced to,

$$\frac{4p(r)}{4p} = \frac{1}{2} \left\{ \exp\left(-\frac{r^2}{L^2}\right) + \exp\left(-\frac{r^2}{M^2}\right) \right\}, \quad (10)$$

$$\begin{aligned} v_{gII} = & -\frac{1}{2}fr + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho} \right. \\ & \times \left\{ \left(\frac{r}{L} \right)^2 \exp\left(-\frac{r^2}{L^2}\right) \right. \\ & \left. \left. + \left(\frac{r}{M} \right)^2 \exp\left(-\frac{r^2}{M^2}\right) \right\} \right]^{\frac{1}{2}}, \quad (11) \end{aligned}$$

and,

$$\begin{aligned} \zeta_{gII} = & -f + \left[\left(\frac{1}{2}fr \right)^2 + \frac{4p}{\rho} \left\{ \left(\frac{r}{L} \right)^2 \exp\left(-\frac{r^2}{L^2}\right) \right. \right. \\ & \left. \left. + \left(\frac{r}{M} \right)^2 \exp\left(-\frac{r^2}{M^2}\right) \right\} \right]^{-\frac{1}{2}} \\ & \times \left[\frac{1}{2}f^2r + \frac{4p}{\rho} \left\{ \left(\frac{r}{L} \right)^2 \left(\frac{2}{r} - \frac{r}{L^2} \right) \right. \right. \\ & \times \exp\left(-\frac{r^2}{L^2}\right) + \left(\frac{r}{M} \right)^2 \left(\frac{2}{r} - \frac{r}{M^2} \right) \\ & \left. \left. \times \exp\left(-\frac{r^2}{M^2}\right) \right\} \right]. \quad (12) \end{aligned}$$

On the other hand, if the motion is assumed to be stationary and axial symmetry, the equation of motion in cylindrical coordinates with frictional layer ranging between a few tens meters and 1 to 2 kilometers above the ground will be given as,

$$\left. \begin{aligned} -(f+\zeta)v_\theta &= -\frac{1}{\rho} \frac{\partial P}{\partial r} + \nu \frac{\partial^2 v_r}{\partial z^2} \\ (f+\zeta)v_r &= \nu \frac{\partial^2 v_\theta}{\partial z^2} \end{aligned} \right\}, \quad (13)$$

where ν : eddy frictional coefficient,

$$P = p + \frac{1}{2}\rho(v_r^2 + v_\theta^2),$$

z : height.

The wind at higher level is assumed to approach to gradient wind, and the average vorticity, averaged from surface to the height of frictional influence is used.

In this case, Dr. Syono has derived the general solution of this equation of motion with frictional layer.

Then Syonos' solutions are,

(a) first case $u=v=0$ at $z=0$, then

$$\left. \begin{aligned} v_r &= -v_g \exp\left(-\frac{\pi}{D}z\right) \sin\left(\frac{\pi}{D}z\right) \\ v_\theta &= v_g \left[1 - \exp\left(-\frac{\pi}{D}z\right) \cos\left(\frac{\pi}{D}z\right)\right] \end{aligned} \right\}, \quad (14)$$

where,

$$\bar{\zeta} = \frac{1}{D} \int_0^{\bar{D}} \zeta dz = 0.827 \zeta_g, \quad (15)$$

$$D = \pi \sqrt{\frac{2\nu}{f + \bar{\zeta}}}, \quad (16)$$

(b) second case, $u = \chi \frac{\partial u}{\partial z}$, $v = \chi \frac{\partial v}{\partial z}$ at $z=0$, then

$$\left. \begin{aligned} v_r &= -v_g C \exp\left(-\frac{\pi}{D}z\right) \cos\left(\frac{\pi}{D}z - \alpha\right) \\ v_\theta &= v_g \left[1 + C \exp\left(-\frac{\pi}{D}z\right) \sin\left(\frac{\pi}{D}z - \alpha\right)\right] \end{aligned} \right\}, \quad (17)$$

where

$$C = \sin \alpha - \cos \alpha \\ = \left[1 + 2\chi \sqrt{\frac{f + \bar{\zeta}}{2\nu}} + \chi^2 \frac{f + \bar{\zeta}}{\nu}\right]^{-\frac{1}{2}}, \quad (18)$$

$$\tan \alpha = \frac{1 + \chi \sqrt{\frac{f + \bar{\zeta}}{2\nu}}}{\chi \sqrt{\frac{f + \bar{\zeta}}{2\nu}}}, \quad (19)$$

$$\bar{\zeta} = \frac{1}{D} \int_0^{\bar{D}} \zeta dz = [1 - 0.173 C (\sin \alpha - \cos \alpha)] \zeta_g. \quad (20)$$

Then, the followings are put in order to abbreviate the calculation,

$$\frac{\Delta p}{n} \sum_{j=1}^n a_j \exp\left(-\frac{r^m}{r_j^m}\right) \\ = X(\Delta p, n, a_j, r_j, m, r), \quad (21)$$

$$\frac{\Delta p}{n} \sum_{j=1}^n a_j m \left(\frac{r}{r_j}\right)^m \exp\left(-\frac{r^m}{r_j^m}\right) \\ = Y(\Delta p, n, a_j, r_j, m, r), \quad (22)$$

$$\frac{\Delta p}{n} \sum_{j=1}^n a_j m \frac{r^{m-1}}{r_j^m} \left\{1 + \frac{m}{2} - \frac{mr^m}{2r_j^m}\right\} \exp\left(-\frac{r^m}{r_j^m}\right) \\ = Z(\Delta p, n, a_j, r_j, m, r). \quad (23)$$

In this case, eqs. (1), (3), (6), (15) and (16) would be deduced to,

$$p(r) = p(\infty) - X, \quad (24)$$

$$v_g = -\frac{1}{2}fr + \left[\left(\frac{1}{2}fr\right)^2 + \frac{Y}{\rho}\right]^{\frac{1}{2}}, \quad (25)$$

$$\zeta_g = -f + \left[\left(\frac{1}{2}fr\right)^2 + \frac{Y}{\rho}\right]^{-\frac{1}{2}} \cdot \left[\frac{1}{2}f^2r + \frac{Z}{\rho}\right], \quad (26)$$

$$\bar{\zeta}_1 = 0.827 \left[-f + \left\{\left(\frac{1}{2}fr\right)^2 + \frac{Y}{\rho}\right\}^{-\frac{1}{2}} \times \left(\frac{1}{2}f^2r + \frac{Z}{\rho}\right)\right], \quad (27)$$

$$D_1 = \sqrt{2\nu} \pi \left[0.173f + 0.827 \left\{\left(\frac{1}{2}fr\right)^2 + \frac{Y}{\rho}\right\}^{-\frac{1}{2}} \times \left(\frac{1}{2}f^2r + \frac{Z}{\rho}\right)\right]^{-\frac{1}{2}}, \quad (28)$$

and eq. (14) is rewritten as,

$$\left. \begin{aligned} v_{rI} &= -v_g \exp\left(-\frac{\pi}{D_1}z\right) \sin\left(\frac{\pi}{D_1}z\right) \\ v_{\theta I} &= v_g \left[1 - \exp\left(-\frac{\pi}{D_1}z\right) \cos\left(\frac{\pi}{D_1}z\right)\right] \end{aligned} \right\}. \quad (29)$$

And α_1 and C_1 are obtained from eqs. (18) and (19) using $\bar{\zeta}_1$ in eq. (27),

$$\begin{aligned} \alpha_1 &= \tan^{-1} \left[1 + \frac{\sqrt{2\nu}}{\chi} (f + \bar{\zeta}_1)^{-\frac{1}{2}}\right] \\ &= \tan^{-1} \left[1 + \frac{\sqrt{2\nu}}{\chi} \left[0.173f + 0.827 \times \left\{\left(\frac{1}{2}fr\right)^2 + \frac{Y}{\rho}\right\}^{-\frac{1}{2}} \cdot \left(\frac{1}{2}f^2r + \frac{Z}{\rho}\right)\right]^{-\frac{1}{2}}\right], \end{aligned} \quad (30)$$

$$C_1 = \sin \alpha_1 - \cos \alpha_1, \quad (31)$$

and so,

$$\bar{\zeta}_{II} = [1 - 0.173 C_1 (\sin \alpha_1 - \cos \alpha_1)] \zeta_g, \quad (32)$$

$$\alpha_{II} = \tan^{-1} \left[1 + \frac{\sqrt{2\nu}}{\chi} (f + \bar{\zeta}_{II})^{-\frac{1}{2}}\right], \quad (33)$$

$$C_{II} = \sin \alpha_{II} - \cos \alpha_{II}, \quad (34)$$

$$D_{II} = \pi \sqrt{2\nu} (f + \bar{\zeta}_{II})^{-\frac{1}{2}}, \quad (35)$$

and so, eq. (17) is rewritten as,

$$\left. \begin{aligned} v_{rII} &= -v_g C_{II} \exp\left(-\frac{\pi}{D_{II}}z\right) \cos\left(\frac{\pi}{D_{II}}z - \alpha_{II}\right) \\ v_{\theta II} &= v_g \left[1 + C_{II} \exp\left(-\frac{\pi}{D_{II}}z\right) \sin\left(\frac{\pi}{D_{II}}z - \alpha_{II}\right)\right] \end{aligned} \right\}. \quad (36)$$

Here, the constants a_j , m , n , r_j , Δp and φ are decided already when the form of typhoons' pressure distribution is decided.

And so, the wind velocity distribution is obtained as the function of the height z and the distance from the center r , by using the equations (29) and (36), and deciding the value of the surface wind condition χ , eddies frictional coefficient ν and the density of air ρ which vary considerably by height.

Taylor decided the value of ν at the layer of height 900 meter, using Debsons' observational data, as follows;

$$\begin{aligned} &6.2 \text{ m}^2 \text{ sec}^{-1} \quad \text{at a gale,} \\ &5.0 \text{ m}^2 \text{ sec}^{-1} \quad \text{at a moderate wind,} \\ &2.8 \text{ m}^2 \text{ sec}^{-1} \quad \text{at a light wind.} \end{aligned}$$

And it was recorded that the value of ν was evaluated as $0.077-0.69 \text{ m}^2 \text{ sec}^{-1}$ at an altitude 100-300 meter on the sea, by Scotiyas' boats

Fig. 1 shows the path of the typhoon June with the date.

Here, the symbol $\bigcirc$ is the centers' position by the aircraft observation, and the symbol $\times$ is one by the synoptic analysis.

Typhoon June developed greatly while she was situated on the sea southeast of Kyushu, and the center pressure showed 909 mb on 8–9th. But then, the centers' pressure depth became shallow, and the region of center was spread gradually.

Fig. 2 shows the relation between the variations of pressures at the center and the outside, and the distance between the outside and the center of this typhoon.

The typhoon is found decaying on the evening of 13th on landing onto the south part of Kyushu.

Fig. 3a shows the depth of centers' pressure Δp and the state of its variation, during 6th in generating to 18th in decaying.

Fig. 3a. The variation of the center pressure depth Δp in typhoon June.

Fig. 3b. The variation of $\frac{\Delta p_{400}}{\Delta p}$, $\frac{\Delta p_{300}}{\Delta p}$, $\frac{\Delta p_{100}}{\Delta p}$.

Fig. 3c. The variation of the constants L , M and N which decide the form of typhoon pressure distribution.

In the most developing period, the value of L was about 500 km and M 30 km, and then the value of M increased gradually after the typhoon struck the South Kyushu.

Then, both values became same magnitude, that is to say, the value of M increased gradually and the both value became to lie one upon another.

Fig. 3b shows the state of variation $\frac{\Delta p(r)}{\Delta p}$, at $r=100$ km, 300 km and 400 km respectively, where

$$\Delta p(r) = p(\infty) - p(r),$$

$$\Delta p = p(\infty) - p_0.$$

p_0 : pressure at the typhoons' center.

It is found that the value of $\frac{\Delta p(r)}{\Delta p}$ was increased, according to that the centers' depth became shallow gradually, after landed onto Kyushu.

In this case, the value of $\frac{\Delta p_{300}}{\Delta p}$ and $\frac{\Delta p_{400}}{\Delta p}$ varied in larger extent than $\frac{\Delta p_{100}}{\Delta p}$.

Fig. 3c shows the changing state of the constants L , M and N , which decide the form of the pressure distribution, property and decline of the typhoon, the pressure distribution of the typhoon being assumed indicated by eq. (7) or (10),

And after, these values increased to about 700 km or 900 km.

Next, the pressure distribution was derived in good approximation applying equation (7) to this typhoon.

The value of N was about 300 km on 11th, then it increased to about 2,000–3,000 km gradually according to the declining.

Fig. 4. The distribution of relative vorticity calculated by eq. (9), in the region of the typhoon June, at 0600z 12th.

Fig. 4 shows the distribution of relative vorticity calculated by eq. (9), in the region of this typhoon at 0600z September 12th.

Then, the large value of relative vorticity, $4 \cdot 10^{-4} \text{ sec}^{-1}$ at $r=100 \text{ km}$, is obtained.

But the value decreases gradually with proceeding to outer region, becomes zero at $r=500 \text{ km}$, and becomes negative and small value in the more outer region.

Here, the following values are used for calculations,

density of air $\rho = 1.1 \cdot 10^{-3} \text{ g cm}^{-3}$,

latitude $\varphi = 27^\circ$

$N = r_1 = 300 \text{ km}$,

$\Delta p = 60 \text{ mb}$.

Fig. 5a. The distribution of calculated wind velocity in the region of typhoon June, as the function of r and z , using the eq. (29), at 0600z 12th.

Here, the values, $a_1=1$, $m=1$, $n=1$, $r_1=300 \text{ km}$, $\varphi=27^\circ$, $\nu=0.3 \text{ m}^2 \text{ sec}^{-1}$ and $z=0 \text{ m}$, 20 m , 50 m are used, and the height of frictional influence is given by eq. (28) too.

Fig. 5a shows the distribution of calculated wind velocity in the region of typhoon June, as the function of r and z , using equation (29), at 0600z September 12th.

Here, the values, $a_1=1$, $m=1$, $n=1$, $r_1=N=300 \text{ km}$, $\varphi=27^\circ$, $\nu=0.3 \text{ m}^2 \text{ sec}^{-1}$ and $z=0 \text{ m}$, 20 m , 50 m are used. And the height of frictional influence is given by eq. (28) too.

Fig. 5b. The distribution of calculated wind velocity and the height of frictional influence by eq. (36) and (35), as the function of r and z , the constants used being same as in Fig. 5a except $x=10 \text{ m}$.

Fig. 5c. The same as Fig. 5b except $x=50 \text{ m}$.

And Figs. 5b and 5c show the calculated wind velocity and the height of frictional influence from eq. (36) and eq. (35) respectively, the same constants and condition as in Fig. 5a being used, except that the value of x is 10 m and 50 m at Fig. 5b and 5c respectively.

The distribution of the observed surfaces' wind velocity as the function of r from the synoptic data, at 0600z 12th, is shown in Fig. 6.

In this case, distinct tendency about the wind velocity distribution is not found owing to the scanty of synoptic data in the inner region of typhoon, though the data in the

Fig. 6. The distribution of observed velocity on sea and on land respectively by synoptic data, at 0600z 12th, in typhoon June.

Symbol $\square$ is data on sea, $\bullet$ stations' data on land, --- the average distributional curve.

outer side are numerous comparably.

In this figure, symbol $\bullet$ gives stations' data on land, and symbol $\square$ on sea.

In Fig. 7a~7e, the distribution of observed wind velocity, as the function of r at surface, in the region of typhoon June is shown.

The data used was selected from only stations in solitary islands and headland, which are not disturbed by topography.

Further, the stations cited are situated on the typhoon center passage or in its neighbourhood.

Fig. 7a. Variation of wind at Kusagakijima during typhoon June passage, in 12-14th.

Symbol $\circ$: data before typhoon inversion, $\bullet$: data after.

Fig. 7a shows wind data at Kusagakijima during typhoon June passage, in 12-14th.

In this figure, symbol $\circ$ gives data before typhoon inversion (1500z 12th-0200z 13th), and symbol $\bullet$ data after (0300z 13th-1500z 14th).

Maximum wind velocity is 36 m sec^{-1} at $r=300 \text{ km}$, and seems closely relating to the calculated gradient wind velocity in Fig. 6c. The agreement can be said good when the height of this station 140.2 meters is considered.

Fig. 7b. The variation of wind velocity at Satamisaki during the typhoon June, the station altitude being 60.2 meters, and the symbol being the same as in Fig. 7a.

Next, wind data at Satamisaki are shown in Fig. 7b, the symbol used being the same as in Fig. 7a.

From the data, the wind velocity distribution is found in fair agreement with the calculated value in Fig. 5c for $z=50 \text{ m}$, referring the station altitude 60.2 meters, and the maximum wind velocity 35 m sec^{-1} at $r=250 \text{ km}$.

Fig. 7c. The variation of wind velocity at Toimisaki during 12th to 14th, the station altitude being 255 meters.

Fig. 7c shows wind data at Toimisaki, during 1500z 12th to 0300z 14th, symbol $\circ$ being data for 1500z 12th-0500z 13th, and $\bullet$ for 0600z 13th-0300z 14th.

Fig. 7d. The variation of wind velocity at Yakushima during 11–14th, the station altitude being 15 meters.

Fig. 7d shows wind data at Yakushima, symbol $\bigcirc$ being data for 0200z 11th–0300z 12th, and $\bullet$ for 0400z 12–0300z 14th.

When the center of typhoon arrived at Yakushima, the wind velocity weakened suddenly and became zero.

After it passed, wind velocity increased sharply again. And the maximum wind velocity attained to 33 m sec^{-1} at $r=250 \text{ km}$, agreeing with Fig. 5b, $z=20$ meters.

Fig. 7e. The variation of wind velocity at Makurazaki during typhoon June, the station altitude being 31 meters.

Fig. 7e shows the wind data at Makurazaki, the station altitude being 30 meters, and the wind velocity resembles to Fig. 6, where $z=20$ meters.

Table 1. Station position and altitude.

Station name	Lat. (N) deg. min. sec.	Long. (E) deg. min. sec.	Altitude (m)
Kusagakijima	$30^{\circ}51'29''$	$129^{\circ}27'56''$	140.2
Satamisaki	$30^{\circ}51'19.5''$	$130^{\circ}39'43''$	60.2
Toimisaki	$31^{\circ}21'49''$	$131^{\circ}20'53''$	255.0
Yakushima	$30^{\circ}27'$	$130^{\circ}30'$	15
Makurazaki	$31^{\circ}16'$	$130^{\circ}18'$	31

Maximum wind velocity is 21 m sec^{-1} at $r=290 \text{ km}$.

Next, the station position and altitude are given in Table 1.

Nextly, Fig. 8 shows the calculated wind velocity on land and sea respectively, in the region of typhoon June, at 0600z 12th, using

Fig. 8. The wind velocity on land and sea respectively, in the region of typhoon June, at 0600z 12th, using Takahashi's conversional graph (Fig. 9a) with the gradient wind velocity.

Takahashi's conversional graph (Fig. 9) with the gradient wind velocity.

In the figure, observational wind velocity on sea is shown larger than conversional value on sea.

Fig. 9a. The relation between the ratio V_s/V_g and the latitude by Dr. Takahashi,

V_s : surface wind velocity,
 V_g : gradient wind velocity.

Fig. 9b. The relation between the angle α , between the isobar and the surface wind, and the latitude.

Fig. 9a is a graph to get wind velocity on sea and land, using the gradient wind velocity by Dr. Takahashi's method.

Fig. 9b is a calculated graph to get the angle between the surface wind direction and isobar as a function of latitude.

Fig. 10. The relation between the wind velocity and the altitude in the region of typhoon June at 0600z 12th, using eq. (36), where the following values are used; $\rho = 1.1 \cdot 10^{-3} \text{ g cm}^{-3}$, $m=1$, $n=1$, $r_1=300 \text{ km}$, $\Delta p=60 \text{ mb}$, $\varphi=27^\circ$, $\nu=0.3 \text{ m}^2 \text{ sec}^{-1}$, $x=50 \text{ m}$.

Symbols $\otimes \cdots \otimes$ show the height of frictional influence.

Fig. 10 shows the relation between the calculated wind velocity and the altitude in the region of typhoon June, at 0600z 12th, using eq. (36), where the following values are used;

$$\rho = 1.1 \cdot 10^{-3} \text{ g cm}^{-3}, \quad m=1, \quad n=1, \\ r_1=300 \text{ km}, \quad \Delta p=60 \text{ mb}, \quad \varphi=27^\circ, \\ \nu=0.3 \text{ m}^2 \text{ sec}^{-1}, \quad x=50 \text{ m}.$$

Here, the longitudinal axis gives the height and the transverse axis wind velocity.

Symbols $\otimes \cdots \otimes$ show the height of frictional influence. From the figure, the height is found very low in the region of this typhoon center.

Fig. 11 shows the vertical distribution of upper wind velocity and the calculated value by eq. (36) at Kagoshima and Itazuki.

Fig. 11. The vertical distribution of wind velocity and the calculated value by eq. (36) at Kagoshima and Itazuki. The explanation for each curve is given in Table 2.

Table 2. Upper wind data and the calculated upper wind velocity at Kagoshima and Itazuki.

	Remarks
A	Kagoshima upper wind data at 0300z 12th, $r=550 \text{ km}$.
B	Itazuki upper wind data at 0900z, 12th, $r=670 \text{ km}$.
C	Values calculated by eq. (36) $x=10 \text{ m}$, $r=800 \text{ km}$.
D	$m=1$, $n=1$, $x=50 \text{ m}$, $r=800 \text{ km}$.
E	$r_1=300 \text{ km}$, $x=10 \text{ m}$, $r=600 \text{ km}$.
F	$\nu=0.3 \text{ m}^2 \text{ sec}^{-1}$, $x=50 \text{ m}$, $r=600 \text{ km}$.

The explanation for each curve is given in Table 2.

Then, the data at Kagoshima is found closely relating to the calculated curves D and F, though the data are scanty in the lower layer from surface to 500 meters.

On the other hand, Itazuki's wind velocity, which was weak at the surface, increased much with height, attaining 26 m sec^{-1} at the altitude 400 m, and weakened with height over 900 m.

This curve is slightly similar to E curve

Fig. 12a. The vertical distribution of upper wind velocity in 13—14th at Itazuki, the wind direction at each height being plotted by arrow mark. The explanation for each curve is given in Table 3.

from these upper data, and it seems that the height of frictional influence is about 500 m at Kagoshima ($r=550$ km) and about 550 m at Itazuki ($r=500$ km).

Fig. 12a shows the vertical distribution of upper wind by rawins data in 13—14th at Itazuki.

Wind direction at each height is plotted by arrow mark, wind direction N↓, E←, S↑, W→, and the explanation for each curve is given in Table 3.

Fig. 12b shows the vertical distribution of

upper wind in 11—14th at Kagoshima, the explanation for each curve being given in Table 4.

Table 4. Kagoshima upper wind data.

Curve	Time and Date	The distance from the typhoon center r	The height of frictional influence
A	1500z 11th	870 km	1,200 m
B	0300z 12th	550 km	500 m
C	2100z 12th	300 km	200—500 m
D	0700z 13th	60 km	—
E	1100z 13th	120 km	0 m
F	1500z 13th	250 km	0 m
G	2100z 13th	530 km	500 m

Table 3. Itazuki upper wind data.

Curve	Time and Date	The distance from the typhoon center r	The height of frictional influence
A	0500z 12th	630 km	400 m
B	0900z 13th	180 km	200 m
C	1500z 13th	70 km	120 m
D	2100z 13th	210 km	80 m
E	0300z 14th	510 km	500 m
F	0900z 14th	820 km	700 m

Symbol.....shows the height of frictional influence in Figs. 12a and 12b.

Fig. 13 shows the relation between the height of frictional influence D and the distance from the center r , using the upper data at Kagoshima and Itazuki, in typhoon June.

Fig. 12b. The vertical distribution of upper wind velocity in 11–14th at Kagoshima. Symbol --- showing the height of frictional influence, and the explanation for each curve being given in Table 4.

Fig. 13. The relation between the height of frictional influence D and the distance from the center r , using the upper data at Kagoshima and Itazuki in typhoon June, symbols $\times$, $\odot$ giving the height of frictional influence at Itazuki and Kagoshima respectively.

Symbols $\times$, $\odot$ give the frictional influence at Itazuki and Kagoshima respectively.

From this figure, it is found that the height of frictional influence is about 0–150 meters in this typhoon center, increasing gradually to outer side, and attaining about 1000–1200 meters in the most outer region.

4. Conclusion and remarks

In this report, the equation of pressure

distribution of exponential type was used, the circulars in the region of typhoon being assumed.

And from it, the relative vorticity, the height of frictional influence, and the distribution of wind velocity were obtained as the function of the height z and the distance from the center r .

And calculated values were compared with the synoptic data, taking typhoon June for an example.

Typhoon June has been observed developing most at far off of south Kyushu.

And when she approached to Kyushu, she shallowed its center depth and declined, and the region of center became spread in a considerable extent.

Then, the pressure distribution was got accurately and simply by eq. (7).

The typhoon June is a good example for the equation of pressure distribution (1), as $a_1=1$, $m=1$ and $n=1$ are suited for synoptic data.

(1) The constants which decide the form of pressure distribution in the region of typhoon changed largely and obviously during the period from the developing to the declining.

(2) The value of relative vorticity showed $4 \cdot 10^{-4} \text{ sec}^{-1}$ at $r=100 \text{ km}$, decreased gradually as proceeding to outer region, became zero at $r=500 \text{ km}$, and showed small negatives in

more outer regions. The vorticity distribution by this method shows the similar tendency as the case of wind distribution.

(3) The eddy frictional coefficient ν , in the region of typhoon, is considered as small value as about $0.3\text{--}0.5\text{ m}^2\text{ sec}^{-1}$, in the lower layer (height $0\text{--}500\text{ m}$).

If the value of ν is assumed to be $50\text{ m}^2\text{ sec}^{-1}$, then the calculated wind velocity in the lower layer becomes too small.

(4) The height of frictional influence in the region of typhoon is considered much lower than that of the general atmospheric condition (height $1000\text{--}1500\text{ m}$).

And it is considered that this height is about $0\text{--}200\text{ m}$ in the typhoon center, increasing gradually to outer sides, and attaining about 1000 meters or more at the most outer sides.

These conclusions ascertain that the vertical distribution of upper wind and wind velocity at solitary islands closely relate to the gradient wind velocity in the region of typhoons center.

In the present paper, the relation between the theoretical value and the observational value of wind velocity is examined, taking typhoon June for an example.

In future, the author will test to apply the method for more examples of typhoons, and will endeavour after the generality of the method.

And further, the author wants to study the relations with the other meteorological elements in the region of the typhoon in detail.

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台風圏内の気象要素の分布 (第1報)

富 高 四 郎
(福岡管区気象台)

台風圏内の海面気圧を円型分布とし、指数項からなる精度のよい一般的な分布式を用いると、傾度風速、渦度分布が求まる。

定常で、偏向力、気圧傾度、摩擦力の項からなり、円座標で表わされた二次元の大気の運動方程式の一般解に、上述の値を代入し、渦粘性係数、地表風の状態の常数値をきめることにより、台風圏内の風の分布を中心距離、高度の函数として、求めることができる。また摩擦高度、上昇速度、渦動性降雨量も求まる。

この算出方法のテストとして、1954年九州に襲来した猛台風第12号について、実地計算を行つた。これから、相対渦度も風の分布から直接求まるものと、同じ程度で、内域は正の値で大きく、外域は負の小さい値を示した。

また渦粘性係数は、台風圏内の下層大気中では、一般に考えられているよりもかなり小さい値 $0.3 \text{ m}^2 \text{ sec}^{-1}$ 程度と仮定してみた。

このとき、台風圏内の各地の地表における風速の水平分布、地勢の影響の小さい海上の孤島観測所の風速変化、および鹿児島、板付の上層風速の垂直分布は、上述の理論による算出値と、かなりよい一致を示した。

今後多くの台風についてのテストにより、この方法の普遍化をはかると共に、さらに上昇速度、渦動性降雨量などの気象要素算出の実用化が重要である。