Dai 20 Gô.

(Nukigaki)

Teikiatu (Cyclone) oyobi Kôkiatu (Anticyclone) no Kôzô ni tuite.

Dai 2.

Syoin, Rigakuhakusi, KOBAYASI-Tatuo.

Kono Ronbun wa "Tenmon oyobi Tikyû-buturigaku Syûhô" 1 no Maki 7 gô to, Doitu-go de kono Hôkoku 1 no Maki 7 gô to ni noseta, onazi Namae no Ronbun no Tuduki de aru. Maeno Ronbun no § 1 niwa, Cyclone ga Ondo no Katamuki no aru Taiki no naka wo susunde yuku tokiwa, sono Migi-usiro ni Ondo no kyûni kawaru Sen ga dekiru koto to, Cyclone no naka no hô e nagarekonde ueni noboru Kûki wa, hutatuno kimatta Sakai no Sen (*Boundary-of-centripetal current*) no aidani aru koto to wo nobete aru. Kono Ronbun no Chart 1 to Chart 2 to wa, zissaino Tenkidu no ueni korera no Sen wo hiita mono de aru. Hutoi Sen ga Ondo no kyûni kawaru Sen de, zissai kwansokusareta *Squall-line* to yoku atte iru. Kusarisen ga *Boundary-of-centripetal-current* de aru. Nidû-sen wa Kwansoku ni yoru *Warm-front* de, tuneni *Centripetal-current* no nakani aru.

§ 2 niwa Anticyclone no nakano, sono Tyûsin ka a mita Nagare no Sen ga keisansite aru. Anticyclone no naka dewa (2)-siki no simesu Atai yorimo tuyoi Kaze wa huki enai. Yueni, Kaze no itiban tuyoi Baai tosite, Tyûsin kara R_1 made (2)-siki no simesu Kaze ga huki, sore no sotoni ikurakano Haba de Kaze ga Tyûsin karano Kyori de kawaranai tokoro ga aru to kangae, kore wo R_1 kara R_2 made to si, sore no soto dewa Kaze wa Tyûsin karano Kyori ga masu to tomoni yowaku naru to sureba, Tyûsin kara R_2 madeno Nagare no Sen wa Fig. 1 no yôni naru. Tugini, Tyûsin no mawarini Kaze no hukanai Bubun ga aru to sureba, Fig. 2 no yôni naru. R_2 no soto de, Cyclone no toki to onazini, Ka e wa Tyûsin karano Kyori ni hanpireisite heru to kangaereba, Anticyclone zentai no nakano Nagare no Sen wa Fig. 3 no yôni naru. Kore ni yotte miruto, Sen EBFG no Nakagawa niwa ue kara orite kita Kûki ga ari, Sotogawa niwa Anticyclone no maeno hô kara nagarete kita Kûki ga aru kara, kono Sen wa Discontinuity-line de aru.

§ 3 niwa Cyclone ga Taiki no naka wo susunde yuku Arisama wo kangaete aru. Cyclone no sitano hôno Bubun niwa utimukino Nagare ga tomonawareru kara Tyûsin kara mita Nagare no Sen wa maeno Ronbun no Fig. 1 no yôni naru. 1000 metre dikakuno Takasa ni naruto utimukino Nagare wa nakunaru kara, Nagare no Sen wa kono Ronbun no Fig. 4 no yôni naru. Sôsureba, Taiki no naka wo wake-susumu Hasira wa Kirikuti ga Maru de nakute tyôdo "Stream-line-form" no Kirikuti ni nite iru. Sikasi sono yokono Men no hô ni susunde yuku. Keredomo, kono Arisama ga zutto ue made tuzuite iru towa kangaerarenai. Ueno hôno ôkuno Bubun dewa, Cyclone wa Nagare no nakano Udu no yôna Arisama de aru to kangaeru no ga tekitô de aru. Soregatameniwa, Aturyoku no Bunpu ga Fig. 5 no yôni naranakutewa naranai. Ima Umi no Men no Takasa de marui Cyclone ga, suiheino Hôkô ni Ondo no Katamuki no aru Taiki no nakani aru to sureba, Takasa ga masu ni sitagatte, Ondo no Katamuki to onazi Muki no Kiatu no Katamuki ga kasanatte, aru Takasa dewa Fig. 5 no Arisama ni naranakutewa naranai. Kono Takysa dewa Cyclone wa Nagare no nakano Udu no Arisama ni naru. Zissaino Baai no ôkuno Zairyô kara, sunawati Nippon no Hen wo tootta ikutumono Cyclone no Tyûsin no ugoku Hayasa to sonotokino Ondo no Katamuki to kara, kono Takasa wo keisansite, itinenno aidano Tukiwake ni suruto, Eig. 7 no yôni naru. Kono Takasa yori nao zutto ue made, Ondo no Katamuki ga aru to sureba, takai Tokoro dewa heikôna Kûki no Nagare ga taihen tuyoku natte, Cyclone wa hayai Nagare no nakani Ikari wo orosite iru Hune no yôna Arisama ni naranakutewa naranai. Kore wa totemo kangaerarenai koto de aru kara, ueni nobeta Takasa kara ueniwa Cyclone no naka dewa Ondo no Katamuki ga nai to kangaeru no ga yoi to omowareru. Sunawati, ueni nobeta Takasa ga, Dimen no Eikyô wo ukete suiheino Muki no Ondo no Katamuki ga dekite iru Tokoro de, soreyori ue wa Dimen kara Ondo no tutawaru no ga osoi ueni, Kûki wa itiniti matawa hutuka ni Tyûsin no mawari wo mawatte simau kara, kimatta Muki ni suiheino Ondo no Katamuki wa nai no de arô. Sôsureba, Fig. 7 de miru Takasa ga iwayuru "Surface-layer" de Huyu niwa 2 kiro kara 3 kiro, Natu niwa 2.5 kiro kara 4 kiro ni naru. Igirisu no Dines wa, iroirono Takasa ni okeru, Aturyoku to Ondo to no Correlation kara, Surface-layer no Takasa wo kimete, Huyu ni 1 kiro kara 2 kiro, Natu ni 2 kiro kara 3 kiro to itte iru. Kore ni kuraberuto, kono Ronbun no Yarikata no hô ga sukosi ôkiku dete iru ga, kore wa Nippon to Igirisu to no Ido no Tigai ga omona Gen'in de arô to omowareru.

No. 20.

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On the Mechanism of Cyclones and Anticyclones.

Part II.[1]

By

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Member of the Institute.

§ 1. Examples of Lines of Discontinuity and Boundaries of Centripetal Current drawn on Weather Charts.

In § 1, Part I, the author has explained that a travelling cyclone generally produces on its right hand side a line of discontinuity, which will become a squall line, if there is sufficient temperature gradient in the direction perpendicular to the path of the centre, and that the air, which is destined to flow into the central part of the cyclone and climb up, lies on a belt between the definite boundary lines—Boundaries of Centripetal Current. The diagram of these lines, the line of discontinuity and the boundary of centripetal current, as well as the lines of flow, are shown in Fig. 1, Part I.

Chart 1 (Part II) shows the positions of these lines on an actual weather map at 10^h 24^{th} September 1918. (Some irregular curvings of the isobars caused by obstacles are smoothed up.)

(1) This is the continuation of the author's previous paper with the same title, published in the "Japanese Journal of Astronomy and Geophysics", Vol. I. No. 7 and in German language in Vol. I, No. 7 of this report, which will be called "Part I" in case of reference.

This chart was drawn in the way as explained below. As the observed winds are more or less affected by local conditions, they are not suited for determining the current of air in a wide area. The gradient wind velocities are calculated from the distances of the consecutive isobars. To get the directions and the magnitudes of the actual wind velocities on the earth's surface, the angle between the isobars and the wind directions on the surface is assumed to be 20° on land and 15° on sea and the ratio of the wind speed on the surface to the gradient wind velocity to be 0.6. (The symbols α and J were used to express these two quantities in Part I.)

We started the calculation in Part I by assuming the radius of the "principal part" or the core and the wind velocity in it. Since it is, however, difficult to discriminate accurately the core part from the surrounding region, it is better to find first the singular point, the point B in Fig. 1, Part I, where the air moves with the same velocity and in the same direction as the centre. The relative velocity of air at any point with respect to the travelling centre is found by composing vectorially the wind velocity with the translational velocity of the centre reversed. Thus, we can draw stream lines relative to the centre, or the lines of flow in this case, connecting the directions of these relative velocities at successive points. Through the singular point we can draw two lines of flow, which are the lines of discontinuity and the boundary of centripetal current.

In Chart 1, isobars are drawn at every two millimetres of mercury. The big arrow shows the direction of the translational motion of the centre, the speed being 14 m per second. The thick full line is the line of discontinuity. The chain line of dots and dashes is the boundary of the centripetal current. The double line is the warm front, the full drawn part being found from the observed data, the dotted part prolonged along the line of flow. The real warm fronts must always lie in the centripetal currents. The

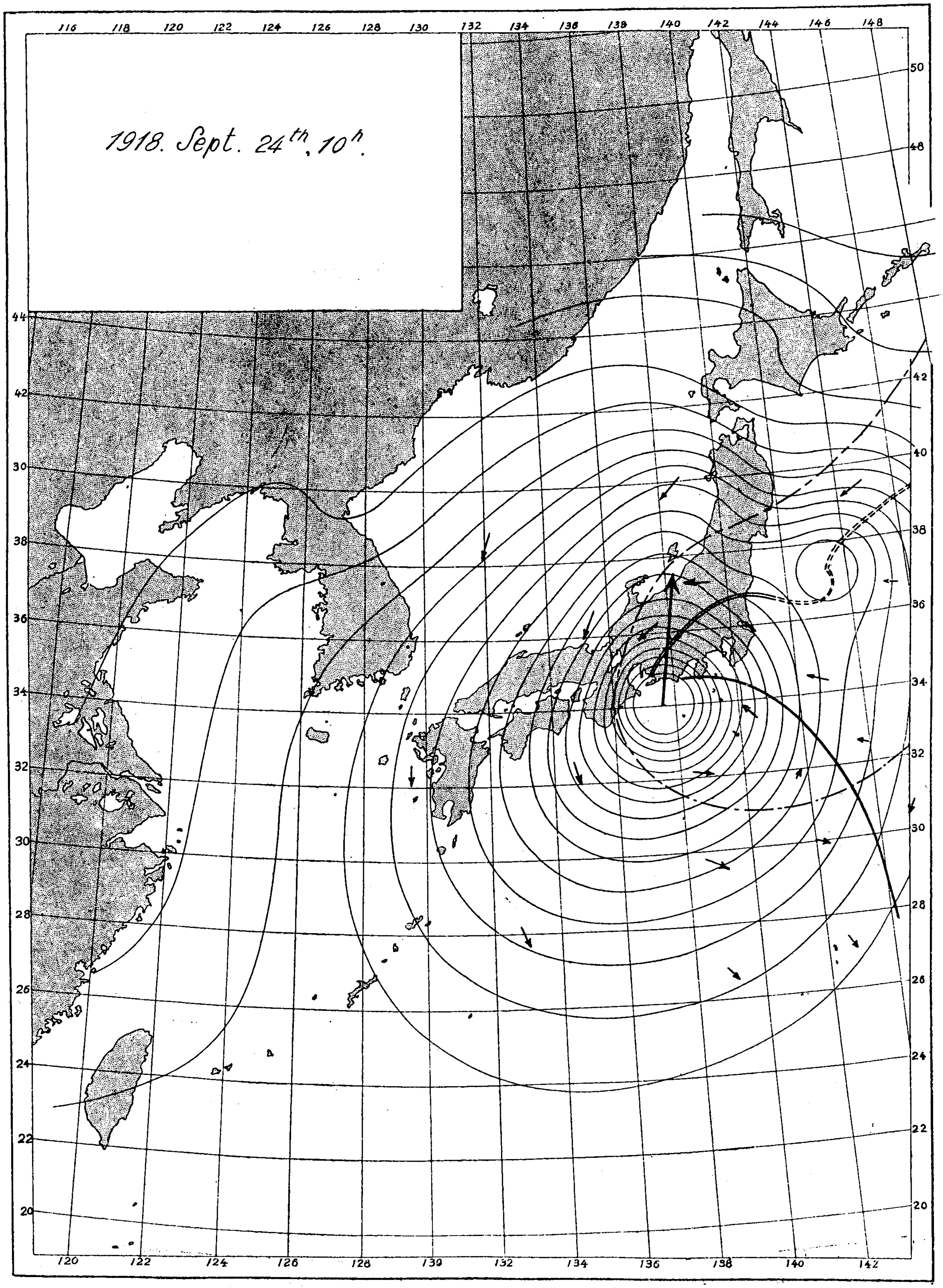

Chart I.

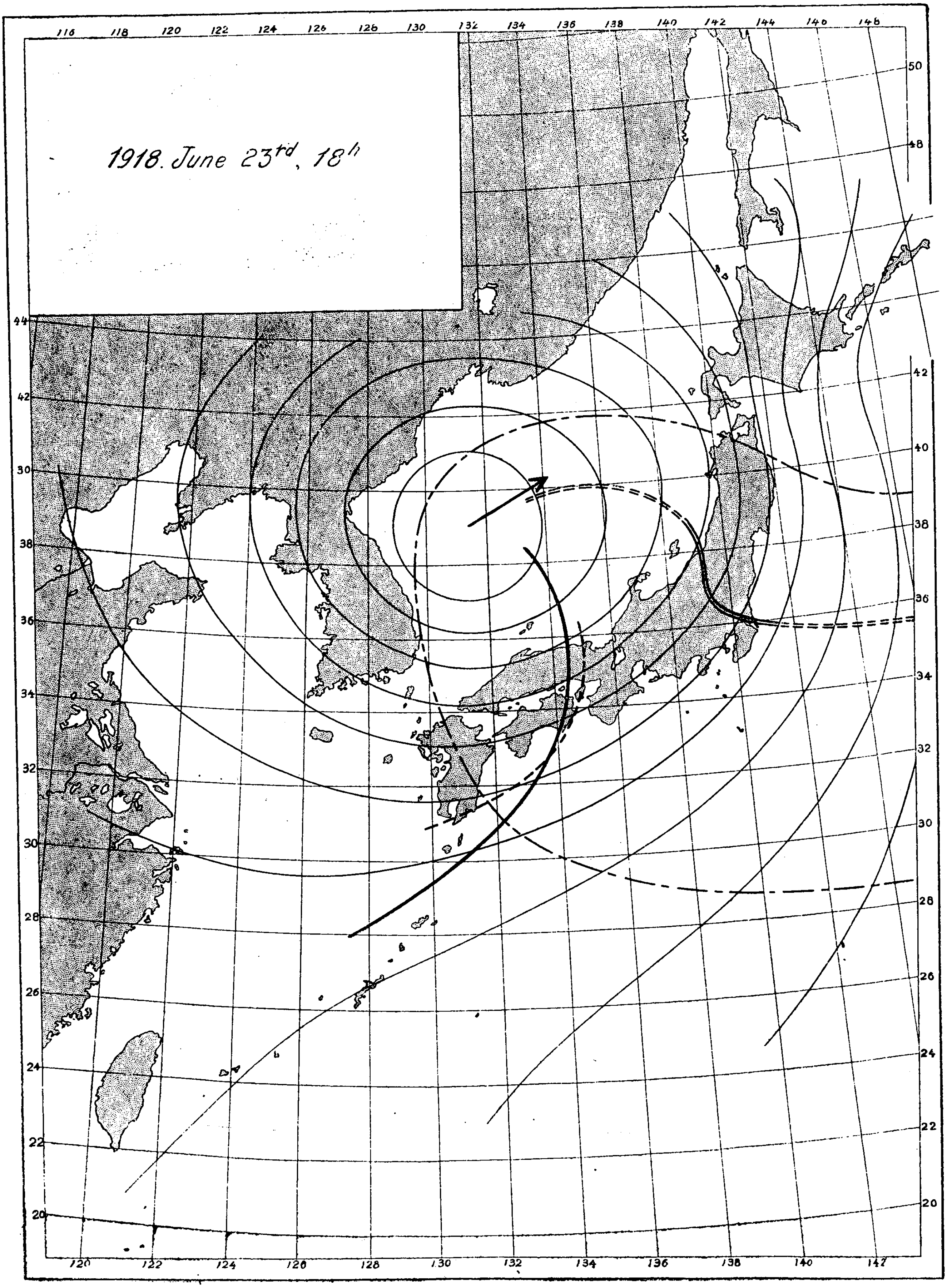

Chart 2.

small arrows shows the directions of the motion of the air relative to the centre. The actually observed squall line just coincides with the line of discontinuity drawn theoretically.

In § 1, Part I, we assumed that the wind velocity in the central region is larger than that of the translation of the centre. If the wind velocity on the right hand side of the travelling centre be smaller than the translational velocity, no line of discontinuity can be produced. In real cases, however, weak cyclones travelling quickly are often accompanied with squall lines. The case of Chart 2 is an example. The gradient wind in the region on the south of the centre within the fifth isobar is 18.5 m per second. If we assume f to be 0.6, the wind velocity on the earth's surface in that region is 11.1 m per second. Since the translational velocity of this cyclone is 13 m per second, it is impossible to develop a discontinuity line on the surface of the earth. But the wind velocity increases very quickly with height in the lower stratum. If f be 0.6 on the surface, it will become about 0.75 at 100 m. Then the wind velocity at that level in the above mentioned region is 13.9 m per second, and we can draw the line of discontinuity and the boundary of the ceutripetal current, as shown in Chart 2.

If the densities of the air on both sides of this discontinuity line be same, the discontinuity (of the history of the air, not of temperature) will remain only on the layer in which it was produced. But, if there be sufficient discontinuity of temperature and the falling down of the cold air across the line (See Fig. 3, Part I and the explanation thereof.) be sufficiently vigorous, the squall line action must reach the surface of the earth.

The thick dotted line in the Chart is the observed squall line. The discontinuity of wind direction was very remarkable along this line, though the temperature discontinuity was not large.

§ 2. Lines of Discontinuity in Anticyclones.

In § 4, Part I, the author remarked that the lines of flow in an anticyclone, except in the central part, have the same feature as Fig. 1, Part I, if the senses of the flow of air and the motion of the centre are reversed, and consequently an anticyclone leaves two lines of discontinuity in its wake. But, as the distribution of the wind velocities in the central part of an anticyclone is entirely different from that in a cyclone, the flow of air in the whole region of the anticyclone will be fully investigated here, under suitable assumptions of wind distribution from the centre to the outmost region.

The equation of the gradient wind in case of anticyclone is

$$\frac{V^2}{r}-2\omega \sin\phi\, V-\frac{1}{\rho}\frac{\partial p}{\partial r}=0,$$

where V is the gradient wind, r the radius of the curvature of the flow of the air, ω the angular velocity of the rotation of the earth, ϕ the latitude, p pressure and ρ the density of the air. Solving this quadratic equation we get

$$V=r\left(\omega\sin\phi-\sqrt{\omega^2\sin^2\phi+\frac{1}{r\rho}\frac{\partial p}{\partial r}}\right). \quad\text{.....................(1)}$$

$\frac{\partial p}{\partial r}$ is negative in case of anticyclone. V is maximum when the expression in the radical is zero, and we get as the greatest possible gradient wind velocity in an anticyclone

$$V=r\,\omega\sin\phi \quad\text{...(2)}$$

If U be the wind velocity on the earth's surface, we can write

$$U=fV. \quad\text{...(3)}$$

Hence in case of maximum wind velocity we get

$$U = Jr\,\omega \sin\phi$$

which we put $= Ar$ ....(4)

If we measure θ from the direction of motion of the centre in counterclockwise, we can put for the rotational velocity

$$\dot{r} = +U\sin\alpha,$$
$$r\dot{\theta} = -U\cos\alpha.$$

When the system is observed from the moving centre, the relative velocity of the general atmospheric field is

$$\dot{r} = -v\cos\theta,$$
$$r\dot{\theta} = +v\sin\theta,$$

where v is the translational velocity of the centre.

Therefore, the whole motion of the air relative to the travelling centre is given by

$$\left.\begin{aligned} \dot{r} &= +U\sin\alpha - v\cos\theta, \\ r\dot{\theta} &= -U\cos\alpha + v\sin\theta. \end{aligned}\right\} \qquad (5)$$

Substituting Ar for U from (4), we can write the equation of stream lines as below.

$$\frac{dr}{d\theta} + \frac{r(Ar\sin\alpha - v\cos\theta)}{Ar\cos\alpha - v\sin\theta} = 0. \qquad (6)$$

The solution of this equation is

$$\tan\alpha\tan^{-1}\frac{Ar\cos\theta - v\sin\alpha}{Ar\sin\theta - v\cos\alpha} - \frac{1}{2}\log\left\{\left(r\sin\theta - \frac{v\cos\alpha}{A}\right)^2 + \left(r\cos\theta - \frac{v\sin\alpha}{A}\right)^2\right\} = \text{const.} \qquad (7)$$

Let us suppose that this condition continues up to a certain distance R_1 from the centre. It will be plausible to assume that there is a zone of more or less breadth,—from R_1 to R_2, say—where the wind speed does not change with the distance from the centre. Denote this constant wind speed in this region by $\overline{U}$. Then $\overline{U}=AR_1$.

In this case the equation of the stream line is

$$\frac{dr}{d\theta}+\frac{r(v\cos\theta-\overline{U}\sin\alpha)}{v\sin\theta-\overline{U}\cos\alpha}=0. \quad \text{(8)}$$

The integral of this equation is

$$\log(\sin\theta-b)+\frac{2\overline{U}\sin\alpha}{v\sqrt{b^2-1}}\tan^{-1}\frac{b\tan\frac{\theta}{2}-1}{\sqrt{b^2-1}}$$
$$+\log r=\text{const.}, \quad \text{(9)}$$

where $b=\dfrac{\overline{U}\cos\alpha}{v}$.

This corresponds to (11) in Part I.

Outside this constant wind zone, the wind velocity must diminish with the distance from the centre. As it was explained in the case of cyclone in § 1 Part I, we may assume

$$Vr=\text{const.} \quad \text{(10)}$$

in the region $r>R_2$.

Since U is continuous at R_2, we get

$$Ur=\overline{U}R_2. \quad \text{(11)}$$

Therefore the equation of the stream line is

$$\frac{dr}{d\theta}+\frac{r(vr\cos\theta-\overline{U}R_2\sin\alpha)}{vr\sin\theta-\overline{U}R_2\cos\alpha}=0. \quad \text{(12)}$$

The solution is

$$vr\sin\theta-\theta\,\overline{U}R_2\sin\alpha-\overline{U}R_2\cos\alpha\log r$$
$$=\text{const.}, \quad \text{(13)}$$

which is different only in sign from (13) in Part I.

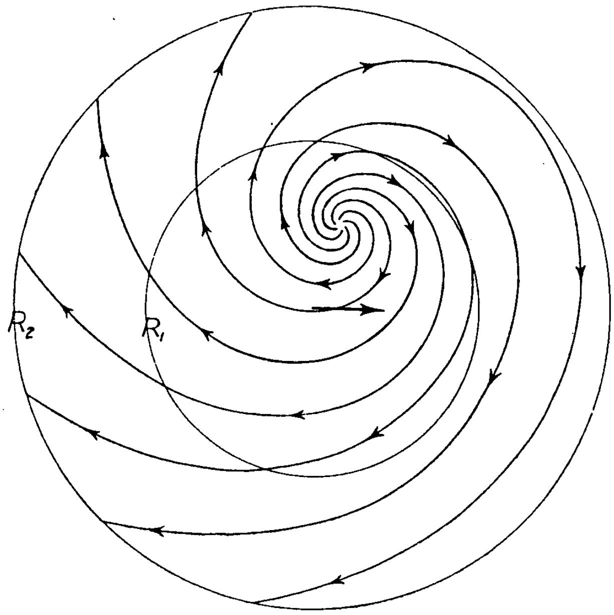

Fig. 1. Lines of flow relative to the centre in the central part of an anticyclone with possibly maximum wind speed.

Fig. 1 shows the lines of flow relative to the centre up to R_2 from the centre, in the case of maximum wind speed in the central part. v is taken to be $\frac{1}{2}\overline{U}$.

There is an apparent diverging centre at

$$\left.\begin{aligned} r &= \frac{v}{f\omega \sin\phi}, \\ \theta &= \frac{\pi}{2} - \alpha. \end{aligned}\right\} \qquad (14)$$

If we assume that an anticyclone has a calm central region, it is obvious that the stream lines relative to the centre in that region are straight and parallel. Fig. 2 is drawn under the assumption that it is calm in the region $r < R_0$; the wind speed increases from zero to a certain value $\overline{U}$, when r increases from R_0 to R_1; between R_1 to R_2 the wind speed is constant.

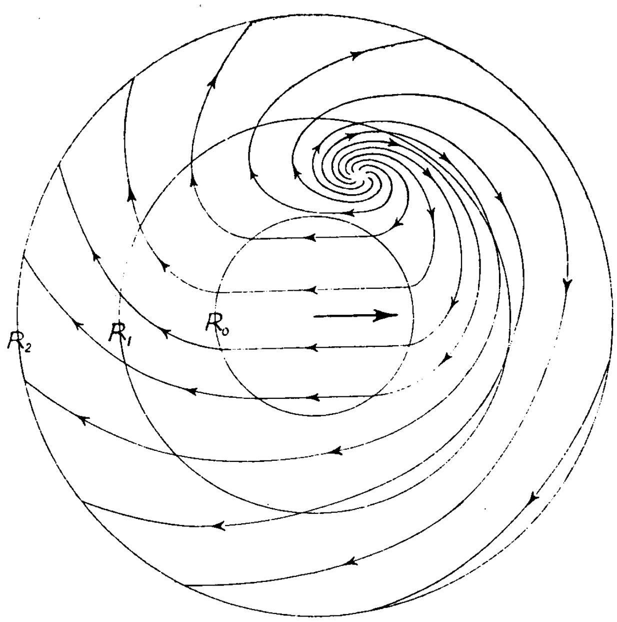

Fig. 2. Lines of flow relative to the centre in the central part of an anticyclone with a calm central region.

Since the above explained two cases are both the extremes, the feature of all anticyclones must lie between these two, provided $\overline{U}$ is larger than the translational velocity. Fig. 3 shows the stream lines in the whole anticyclonic area. As it was mentioned in Part I, these stream lines are at the same time lines of flow or the orbits of air parts seen from the centre. Therefore we see in this diagram that the line EBFG is the boundary of two different systems of flows; one which comes from the front of the anticyclone, (speaking relatively to the moving system) running on the surface of the earth and the other which is of the air descending from the higher level. The temperature of the air in anticyclones is very much affected by radiation, because of the slow motion and the clear sky. Consequently, the jump of temperature at this line of discontinuity is generally small. However, the air on both sides of this line have the different histories, and we often observe the marked discontinuities of water contents. The line ABC is the

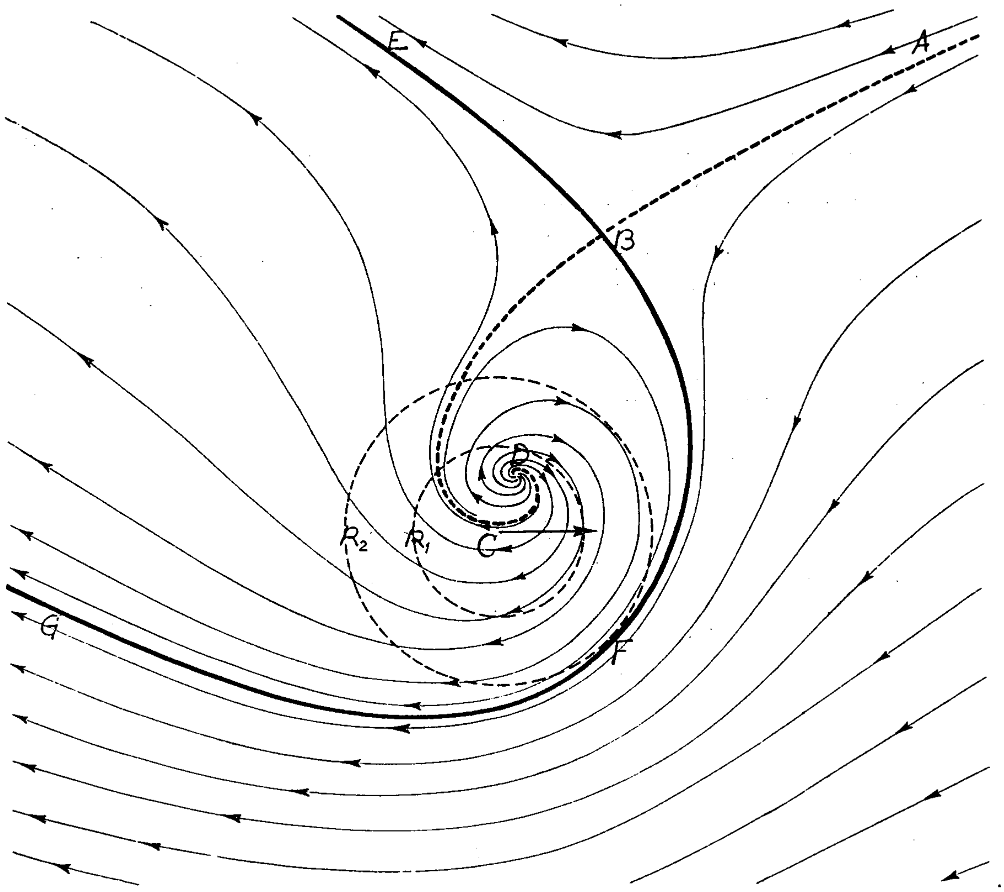

Fig. 3. Lines of flow relative to the centre in a whole an'icyclonic are a.

boundary of the two systems of flows which separate to north and south afterwards. Therefore, the temperatures on the outsides of the both discontinuity lines (on the northern side of the northern discontinuity line and on the southern side of the southern discontinuity line) will be equal, if they are not affected by the earth's surface. For the inclinat'on of the discontinuity surface, or the variation of the positions of the discontinuity lines at different levels, see Fig. 3 in Part I, turning upside down.

§ 3. On the Translational Motion of Cyclones and Anticyclones.

In § 1, Part I, we calculated the lines of flow in the moving cyclone with inward flow. They are shown in Fig. 1, Part I. If we assume that at a certain height, say 1000 m, the inward component of the wind vanishes and the wind blows along the isobars, we can find the lines of flow (with respect to the travelling system) above that height, putting

$$\alpha=0, \quad J=1$$

in the expressions (11) and (13) in Part I.

Or, we can calculate them independently as follows.

V be the velocity of the gradient wind and v the translational velocity of the cyclonic system as before. Since V is a function of r only, we can put

$$r\dot{\theta}=V=f(r).$$

Then the motion of air relative to the centre can be written

$$r\dot{\theta}=f(r)+v\sin\theta,$$

$$\dot{r}=-v\cos\theta.$$

Hence the equation of the stream line is

$$\frac{d\theta}{dr}=-\frac{f(r)+v\sin\theta}{rv\cos\theta} \qquad \text{.......................................(15)}$$

The solution is

$$\int f(r)\,dr+rv\sin\theta=\text{const.} \qquad \text{..............................(16)}$$

If we assume the law of wind distribution outside the core region to be

$$Vr=\text{const.},$$

(16) becomes

$$C \log r + rv \sin \theta = \text{const.}$$

Therefore, the line of flow relative to the centre above 1000 m level is as shown in Fig. 4. We see in this diagram that the horizontal section of the travelling column is not circular, (when the pressure distribution at the sea level is circular); it is very like a "stream line from section," but it travels sideways.

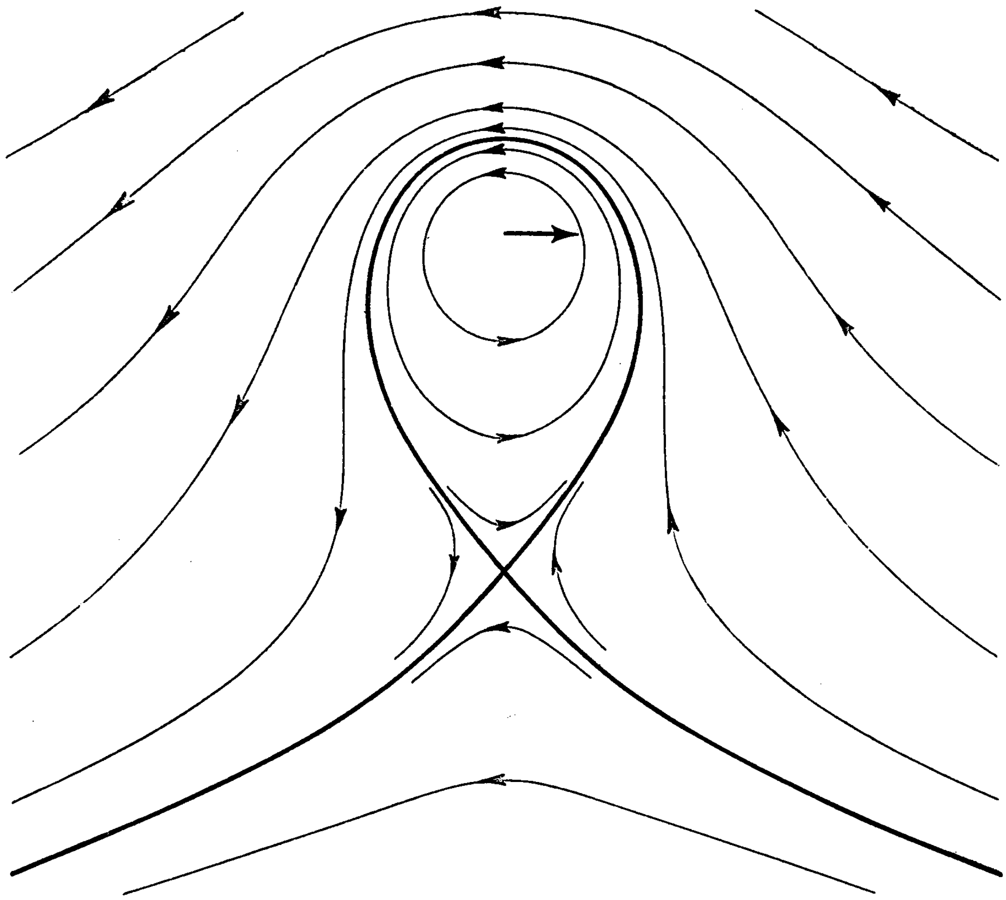

Fig. 4 Lines of flow relative to the centre at 1000 metre level in a circular cyclone.

It does not seem likely that this condition continues up to a great height and such a column breaks through the general atmosphere. It is much more plausible to regard the cyclonic motion in the upper atmosphere as a vortex in stream.

If it be so, the pressure distribution in such a layer must be the superposition of parallel (straight isobars parallel to the direction of the translational motion) and radial (circular concentric isobars) gradients; the former corresponding to the general current of the atmosphere, the latter to the revolving motion. Then, all the trajectories or the paths of air seen from the earth's surface are trochoids. Fig. 5 is an example of the pressure distribution of this kind. In this example, the translational velocity is taken to be 10 m per second to the Eastward, the centre being at 40° latitude, and the revolving speed in the central region (shown by the dotted circle) to be 30 m per second. The isobars are drawn for every millimetre of mercury. The density of air used in this calculation corresponds to the 3000 m level.

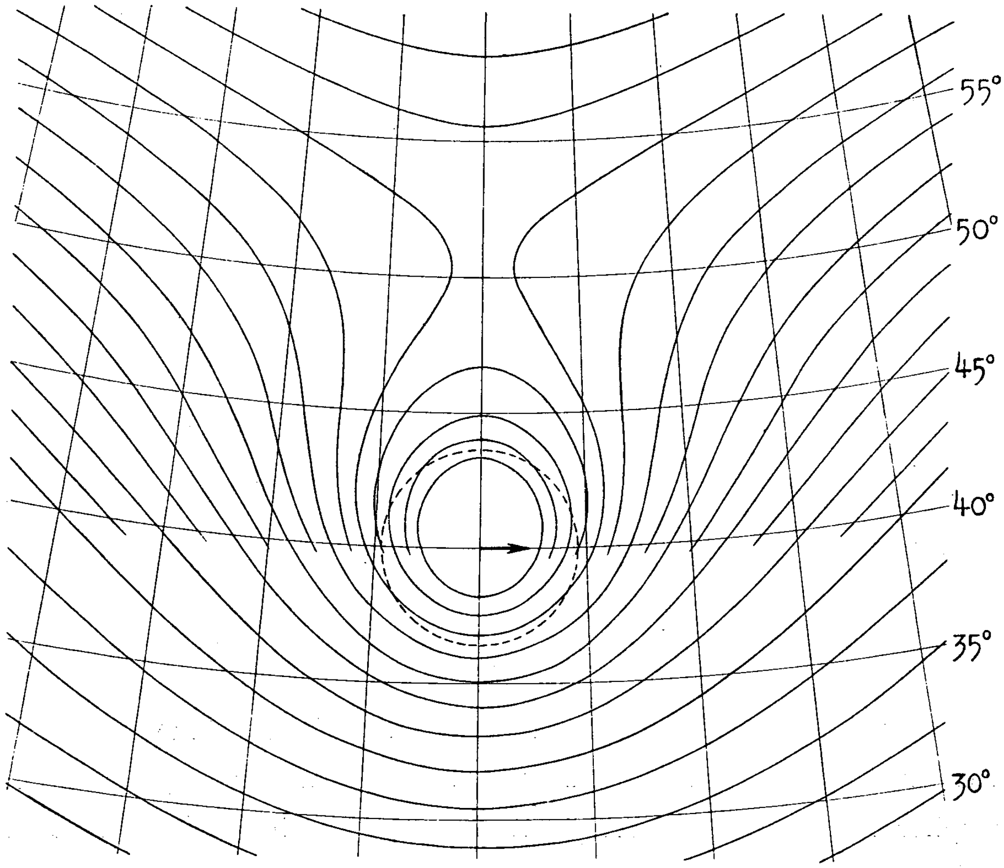

Fig. 5. Pressure distribution at 3000 metre level in a cyclone which is circular at the sea level.

Let us assume that the isobars at the sea level are circular and concentric. If there is a horizontal temperature gradient in a definite direction in the whole area considered, the superposing parallel pressure gradient in the same direction as the temperature gradient must exist in a higher level. This pressure gradient increases with height and at a certain level the pressure distribution of the form of Fig. 5 will be reached. The height of such a level, or the height at which the circular cyclone acquires the vortex-in-stream character, will be calculated in the following.

Napier Shaw(1) has deduced the expressions, which give the vertical changes of the pressure gradient and wind velocity in tems of pressure, temperature and their horizontal gradients.

Let us find the integral form of the similar equation, which is covenient for our purpose.

Since the dynamical pressure change in vertical direction is very small, we may start from the statical equation

$$\frac{\partial p}{\partial h}=-g\rho$$

where p is the pressure, h the height from the sea level, g the gravitational acceleration and ρ the density of the air.

If we assume the vertical lapse rate of temperature λ to be constant, or if we take the mean value of it between the surface and the point considered, we can write

$$\rho=\frac{\rho_0 T_0 p}{p_0 T}=\frac{\rho_0 T_0 p}{p_0(T_s-\lambda h)}. \qquad (18)$$

where T is the absolute temperature and ρ_0, p_0, T_0 are ρ, p, T in the standard condition respectively, and T_s is T at the sea level.

Therefore (17) becomes

$$\frac{\partial p}{\partial h}=-\frac{gT_0\rho_0}{p_0}\frac{p}{T_s-\lambda h}. \qquad (19)$$

(1) See 'Manual of Meteo:ology. Part IV.' p 79.

x and y be the rectangular coordinates in the horizontal plane then the general solution of this partial differential equation is

$$\frac{p^{c\lambda}}{T_s-\lambda h}=\Phi(x,y), \qquad \text{(20)}$$

where Φ is an arbitrary function, and

$$c=\frac{p_0}{T_0\rho_0 g}. \qquad \text{(21)}$$

At the sea level, (20) becomes

$$\frac{p_s^{c\lambda}}{T_s}=\Phi(x,y). \qquad \text{(22)}$$

where p_s is p at the sea level.

Therefore, we can write

$$\frac{p^{c\lambda}}{T_s-\lambda h}=\frac{p_s^{c\lambda}}{T_s}. \qquad \text{(23)}$$

Differentiating the both sides of (23) with respect to y, we get

$$\frac{c}{p}\frac{\partial p}{\partial y}-\frac{c}{p_s}\frac{\partial p_s}{\partial y}=\frac{h}{T_s(T_s-\lambda h)}\frac{\partial T_s}{\partial y}. \qquad \text{(24)}$$

V_x be the x-component of the geostrophic wind, we can put

$$V_x=\frac{-1}{2\omega\sin\phi\,\rho}\frac{\partial p}{\partial y}=\frac{-1}{2\omega\sin\phi}\frac{p_0 T}{\rho_0 T_0 p}\frac{\partial p}{\partial y}=\frac{-cg}{2\omega\sin\phi}\frac{T}{p}\frac{\partial p}{\partial y}, \qquad \text{(25)}$$

where c comes from (21).

From (24) and (25) we get

$$T_s V_x-(T_s-\lambda h)V_{sx}=\frac{-gh}{2\omega\sin\phi}\frac{\partial T_s}{\partial y}, \qquad \text{(26)}$$

where V_{sx} is the x-component of the undisturbed geostrophic wind at the sea level. This can be written

$$A_x-V_{sx}=\frac{-g}{2\omega\sin\phi}\frac{h}{T_s}\frac{\partial T_s}{\partial y}-\frac{\lambda h}{T_s}V_{sx} \qquad \text{(26)}'$$

This gives the difference of the velocities of the geostrophic winds at the height of h and at the sea level.

Or we can write

$$V_x - \frac{T_s - \lambda h}{T_s} V_{sx} = \frac{-g}{2\omega \sin \phi} \frac{h}{T_s} \frac{\partial T_s}{\partial y}. \quad \text{.......................(26)}''$$

If we take y-axis in the reversed direction of the temperature gradient, (toward north) we have for the other component

$$V_y = \frac{T_s - \lambda h}{T_s} V_{sy}. \quad \text{..(27)}$$

$\frac{T_s - \lambda h}{T_s} V_{sx}$ is the geostrophic wind at the height of h, calculated from the pressure gradient at the sea level, ignoring the horizontal temperature gradient. Therefore, (26)″ gives the superposed parallel straight current in x-direction. If the temperature gradient $\frac{\partial T_s}{\partial y}$ be constant in a whole cyclonic area, (In case the temperature gradient is not uniform, the cyclone becomes distorted; See p. 231) the cyclone reaches the vortex-in-stream state at the level of such a height that the right side of (26)″ becomes equall to the translational velocity. We are going to calculate this height (Let us denote this height by $\boldsymbol{h}$.) from the observed values of translational velocities, temperatures and temperature gradients. The formula to be used is (26)″, which can be written

$$v = \frac{g}{2\omega \sin \phi} \frac{\boldsymbol{h}}{T_s} \frac{dT_s}{dn}. \quad \text{.......................................(28)}$$

where v is the translational velocity and n the direction normal to v.

Let us examine the mean data first. In and in the neighbourhood of Japan, the mean temperature gradient in January is about 1.66° per 100 km from NNW to SSE, the mean travelling velocity of cyclones in Winter is 12.4 m per second and the mean temperature in the whole region is roughly *O° C.* (For the sake of

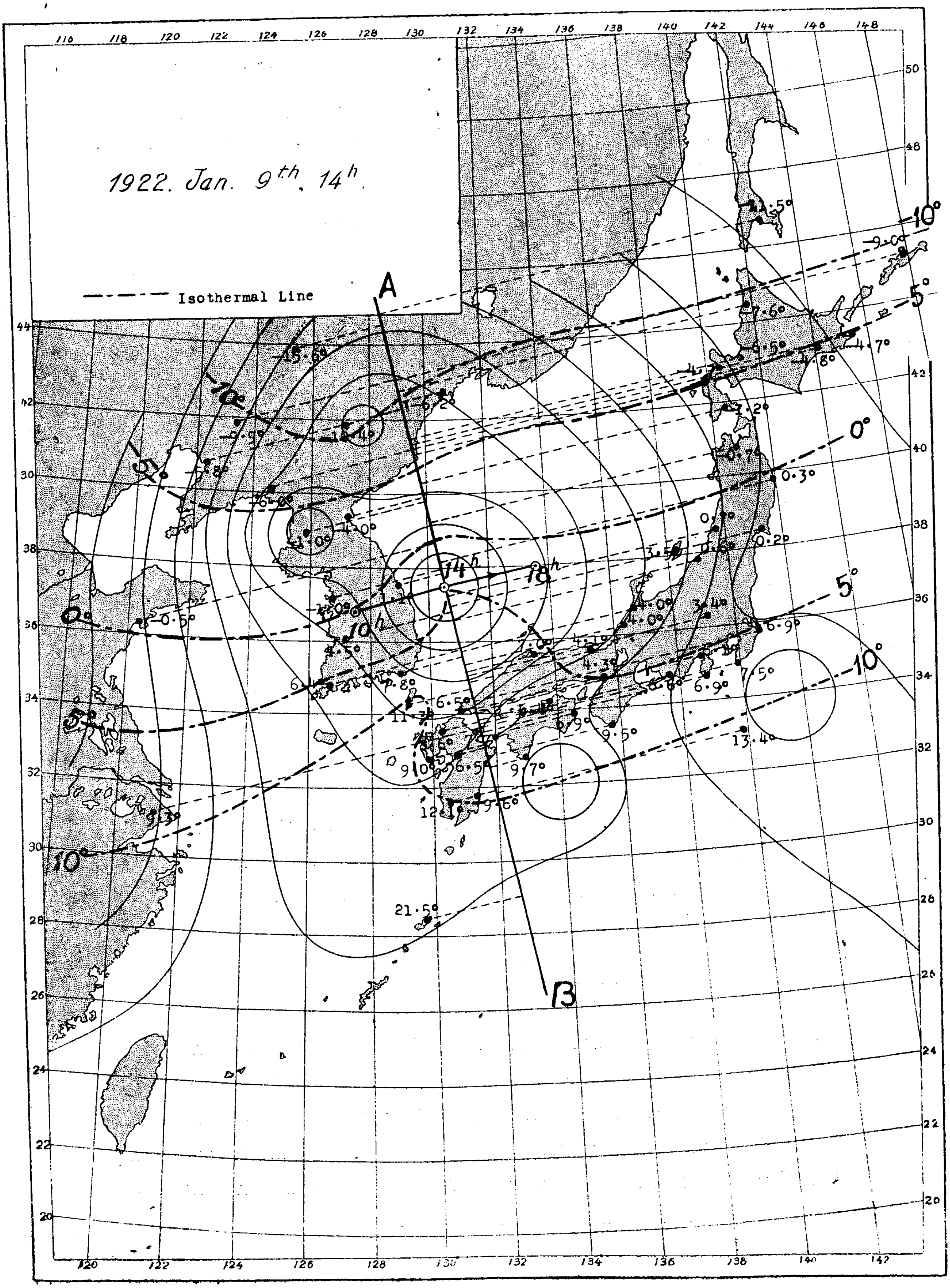

Chart 3.

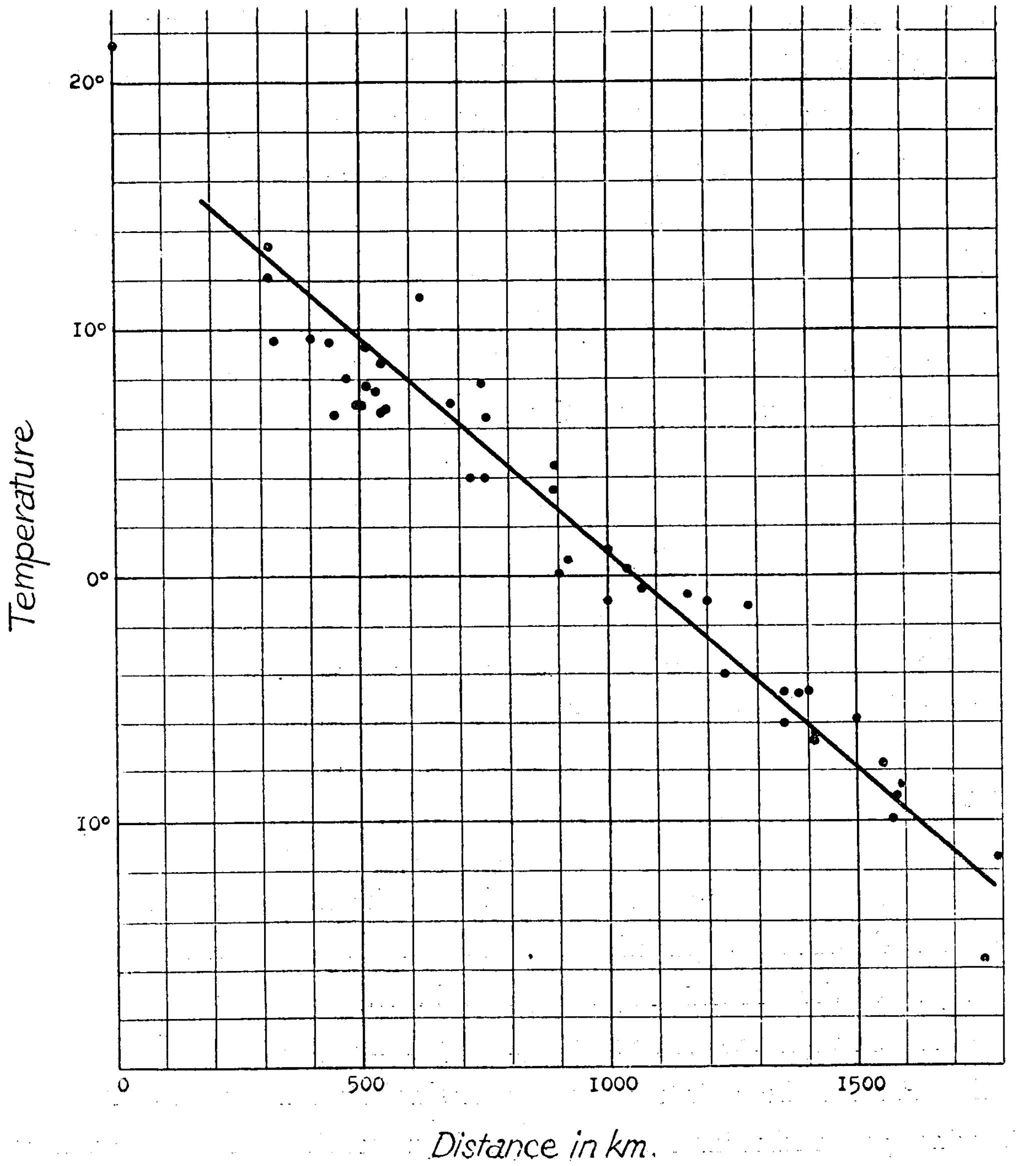

Fig 6. Temperature distribution in a cyclonic area.

simplicty, the variation of T_s, the absolute temperature on the surface, may be neglected in discussing the general current in whole cyclonic area, because the estimation of the other quantities cannot be very accurate; especially the fact described on p. 230 line 22 is not considered here.) In August, they are 0.74° from *N* to *S*, 7.8 m per second and 23° *C.* respectively. (The temperature gradient suddenly changes at the latitude 34° in August; in the whole area on the North side of that latitude the temperature gradient has the above value, while in South it is very small and indeterminate. Therefore the mean translational velocity 7.8 m is perhaps too small for the Northern part.)

Calculating from these data, we get the following values.

$h = 1.95$ km in January,

$\prime\prime = 2.98$ km in August.

Next, we must investigate the individual cyclones. To do this, three successive weather maps with intervals of four hours are made, when a cyclone is travelling across our region. The translational velocity is found from the positions of the centre in the first and the third maps. Isothermal lines are drawn on the second map. Then we always find that the mean direction of the isothermal lines is parallel to the direction of the path. An example of the second maps is given in Chart 3. Draw a straight line AB through the centre of the cyclone perpendicular to the path. Project the points of the observing stations to the line AB, drawing perpendicular lines through these points. In this procedure the curvature of the earth's surface is ignored, as it is not necessary to adopt a very accurate and laborious method in our case. Mark the points on a coordinate paper, taking the positions of the projections of the observation points on the line AB in abscissa, the temperatures at those points in ordinate, as shown in Fig. 6. Then connecting the mean positions of the points, we can draw a line which gives the gradient of temperature.

In the example of Fig. 6, the translational velocity is 19.6 m per second, the temperature gradient 1.75° per 100 km and the mean temperature in whole cyclonic area about 0° *C.* Hence we get

$$\boldsymbol{h} = 2.91 \text{ km}$$

In this way, six cyclones were investigated and the following values of ***h*** were found.

Date				Transl. vel. in m/sec.	Temp. grad. per 100 km.	Mean temp.	***h*** in km.
14^h	9^{th}	Jan.	1922	19.6	1.75°	0°	2.91
14	24	Feb.	22	23.2	2.16°	10°	2.90
14	7	Apr.	18	16.9	1.51°	10°	3.02
10	23	June	18	12.8	1.22°	20°	2.94
6	30	Aug.	18	11.7	0.99°	25°	3.37
6	15	Sept.	18	15.8	1.11°	20°	4.00

In the investigation of the following nineteen cyclones, the labour of drawing special weather maps was avoided, and the daily weather maps at 6^h and 18^h published by the Central Meteorological Observatory of Japan were used. Therefore, the translational velocities given below are the mean velocities in 24 hours.

Date				Transl. vel. in m/sec.	Temp. grad. per 100 km.	Mean temp.	***h*** in km.
18^h	8^{th}	Jan.	1926	12.4	1.25°	3°	2.62
6	4	Feb.	26	16.9	1.82°	5°	2.44
18	3	Feb.	26	15.8	1.77°	6°	2.39
18	13	Mar.	26	16.3	1.52°	10°	2.90
6	14	Mar.	25	18.8	1.77°	5°	2.82
6	3	Apr.	26	16.6	1.62°	10°	2.78
6	21	Apr.	26	11.5	1.16°	10°	2.70
18	14	May	25	10.1	0.78°	15°	3.56
6	30	July	26	12.1	0.96°	18°	3.56
18	3	Aug.	26	16.0	1.60°	22°	2.81
6	30	Aug.	25	12.5	1.26°	20°	2.78

Date				Tran l. vel. in m/sec.	Temp. grad. per 100 km.	Mean temp.	h in km.
6h	4th	Sept.	1926	12.5	1.33°	22°	2.65
6	7	Sept.	25	12.8	0.96°	22°	3.75
18	17	Sept.	26	12.5	0.92°	20°	3.80
18	11	Nov.	25	17.8	2.16°	10°	2.23
6	29	Nov.	25	11.9	1.29°	8°	2.40
18	2	Dec.	25	11.9	1.20°	8°	2.68
18	15	Dec.	25	20.2	2.33°	6°	2.30
6	21	Dec.	25	12.9	1.66°	7°	2.08

These 25 values of h are plotted in Fig. 7, taking the dates in the abscissa. These points seem to lie on a belt between two

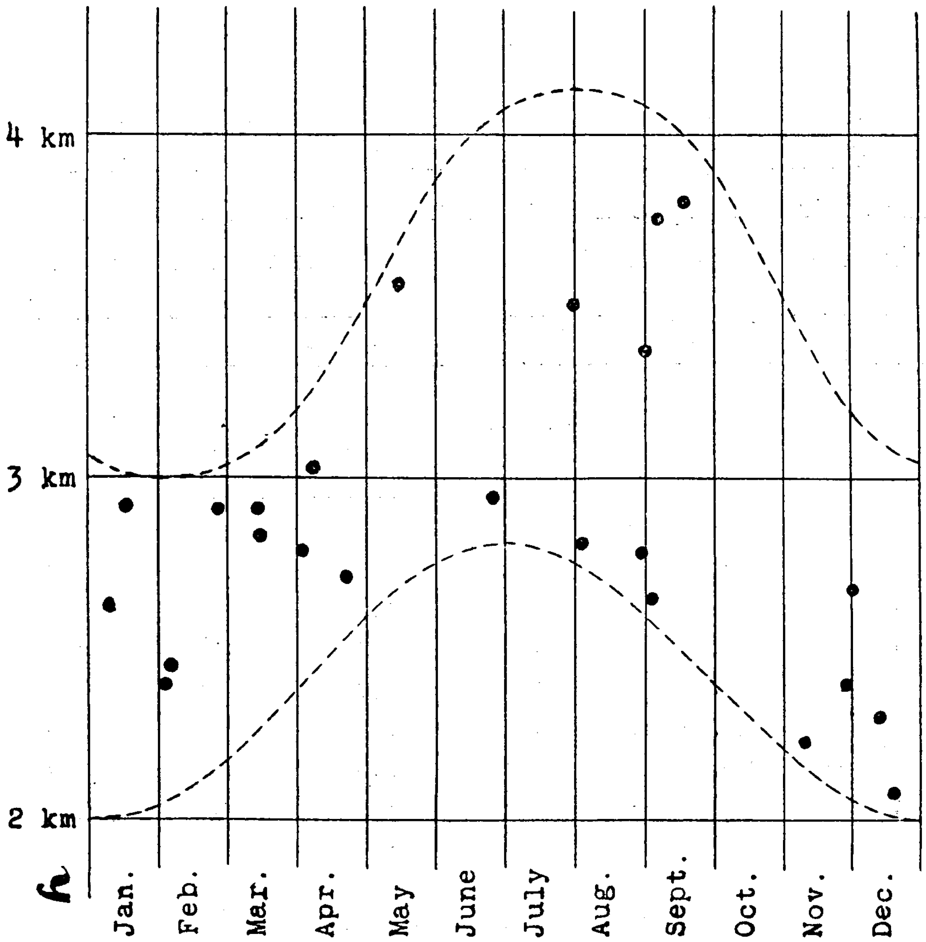

Fig. 7. Variation of h in the year.

sine-shaped curves as shown in Fig. 7, though the curves drawn are somewhat arbitrary. In chosing these examples, all the cyclones which passed our region in certain periods were taken. Therefore, they were observed in various parts of Japan; some were on the sea and some were crossing high muntainous districts at the time they were investigated. If we select the examples only among cyclones which are travelling on similar surface, the width of the belt in fig. 7 will become still narrower.

Now we see that the cyclones acquire the vortex-in-stream character at a height betwen 2.5 and 4 kilometres in Summer and 2 and 3 kilometres in Winter. We may consider these heights as the limit of the layer influenced by the temperature on the surface of the earth. In the layer of the lowest several hundred metres, the air is not conserved in its posittion in the revolving column, but it climbs up in the central part and flows away to the backward of the cyclone in the surrounding region, fresh air constantly flowing in from the front. (See Fig. 1 and 2, Part I.) Therefore, the temperature gradient in the atmospheric field, in which the cyclone is travelling, remains to exist in the cyclonic area, the isothermals curving northwards. This temperature gradient in the direction perpendicular to the translational velocity reaches up to the ***h***-level by the disturbance due to turbulence and convectional motion. Therefore, in the layer below ***h*** there must be horizontal temperature gradient. But, in the layer above ***h***, we may consider that there is no temperature gradient, because the temperature in this layer is affected very slowly by the temperature of the earth's surface, since the turbulence and convection (of ordinary strength) do not reach this height, while the greater part of the air completes its revolution around the centre in one or two days. If, on the contrary, the temperature gradient continue to exist in the same direction up to a still greater height, the parallel current in such a layer must be much greater than the translational velocity of the cyclonic system. It is difficult to think that a cyclone is in the vortex-in-stream state in the neighbourhood of the ***h***-level only, and in the higher layers it is in a state like that of a ship anchored in a stream. (The author asserts the absence of temperture gradient in definite direction above the height of 3 or 4 kilometres only in quickly revolving systems. In other cases, it is well known fact that the wind velocities increase often even in still higher layers.)

W. H. Dines,(1) calculating the correlations between the deviations of temperature and pressure from the normal values in the upper air, determined the height of the "surface layer" by the reason that the correlation coefficient in that layer is very small, while in all layers above, up to nine kilometres, the coefficients are large and nearly equal to one another. His values of the height of the surface layer are between 1 and 2 kilometres in Winter and 2 and 3 kilometres in Summer. Our values of ***h***—2 to 3 kilometres in Winter, 2.5 to 4 kilometres in Summer—are about 70 and 30 per cent. larger than Dines' values. This comes probably from the difference of the latitudes of England and our country. Moreover, since we are considering cyclonic areas only, the climbing current of the air also will make ***h*** larger.

In the above calculation, we regarded the pressure distribution in cyclones at the sea level to be symmetrical around the centre, or considered that the gradient wind on the surface is accompanied by no parallel current. But, in reality, every cyclone has more or less asymmetry in pressure gradient (at the sea level) on both sides of the path. If the isobars run closer to each other on its right hand side, say, than on the left, it may be considered that a parallel current in the same direction as the translational motion of the centre is superposed on the revolving motion of the system,

(1) *Geophysical Memoirs*, No. 13, M.O. publication No. 220*c*.

even on the surface of the earth. In such a case, the vortex-in-stream state is reached at a still lower level than ***h***, calculated by ignoring the asymmetry. This is better to be expressed in the reversed order as follows. Above a certain height ***h***, there is no definite temperature gradient in any horizontal direction in a cyclone, because the same air revolves round the centre quickly. The temperature gradient on the earth's surface extends up to ***h***-level, by turbulent and convectional motion of the air. The pressure distribution at that level, which is of the kind of Fig. 5, is transmitted to the surface of the earth by the layer of air with the horizontal temperature gradient. The effect of this layer on the pressure distribution at the sea level is, in oridinary cases, such as to compensate the asymmetry of the pressure distribution in the higher layers. Therefore, if the temperature gradient is too small to complete the compensation, the pressure gradient at the sea level is greater on the right hand side of the path than on the left, and vice versa.

In calculating the temperature gradient by the method of Fig. 6, we find that the line representing the temperature distribution, which is always very nearly straight in the central and northern parts of the cyclone, often, especially often in Summer, curves in the southern part, and the gradient diminishes there considerably. (This is the case also with the mean temperature in Summer, as mentioned before.) In such a case, since the value of ***h*** is nearly constant in the whole cyclonic area according to the season, we may consider that the superposing parallel current in the upper atmosphere in the direction of the translation of the centre is very much smaller in the southern region than in the central and northern regions. Therefore, the southern part of the cyclone must lag behind the central and northern parts. Fig. 8 shows the feature of the change in the forms of isobars, which were circular at first, under such kind of distribution of the surface temperature. In this figure, the distance in the direction perpendicular to the path of the centre is common to both the isobaric diagram and the curves of the temperature and the velocity. The velocities of the parallel

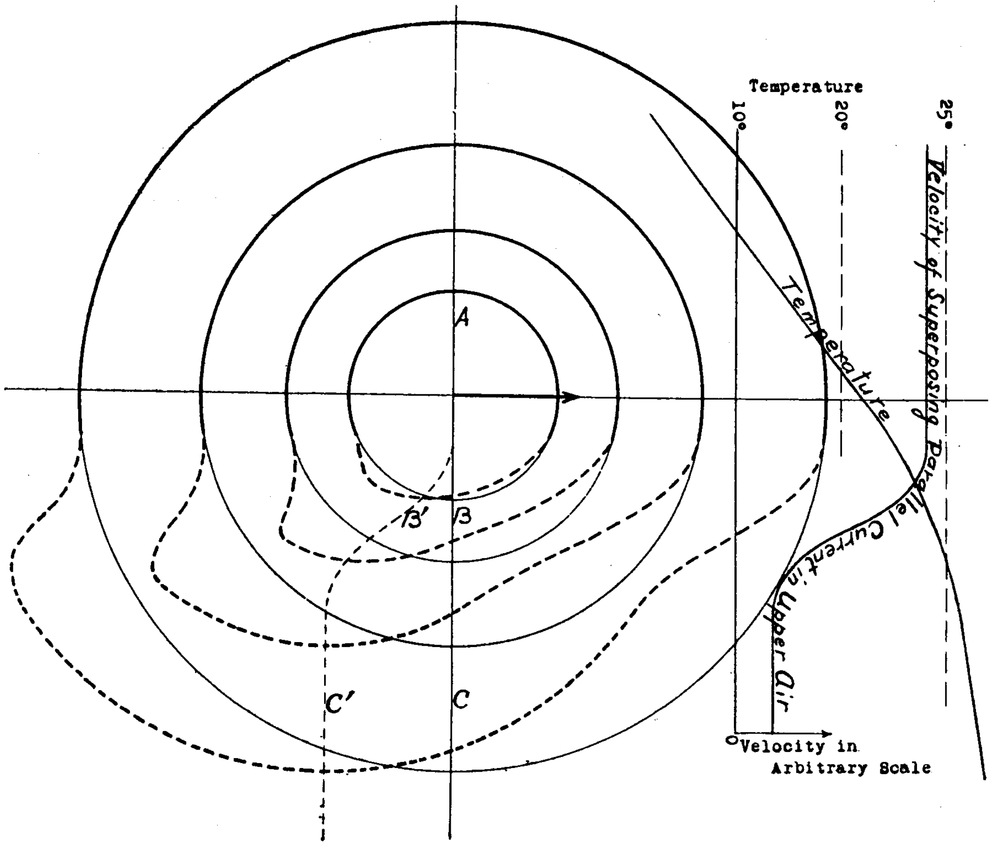

Fig. 8. Distortion of the originally circular cyclone in a region with ununitorm temperatu e gradient.

current in the upper atmosphere were calculated from the assumed temperature gradients, considering ***h*** to be constant, and are given by the curves in arbitrary scale. The trough line ABC curves as AB′C′ after a certain time. Then the isobars become as shown by dotted lines. We often see cyclones which have long trough of the form very like this figure in their rear right. Those troughs may also be produced by the influence of squall lines. But the above mentioned cause is equally possible.

The same calculations were made about anticyclones, and the values of h of the same order as in cyclones were found. However, anticyclones in our region change their forms constantly, and it is difficult to determine the exact values of translational velocities. The distributions of temperatures are also less regular in these cases than in cyclonic cases. This fact seems to be quite natural, because the flows in anticyclones are as shown in Fig. 3.

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