A GENERAL SURVEY OF FACTORS INFLUENCING DEVELOPMENT AT SEA LEVEL

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ABSTRACT

Sutcliffe's theory of development is reviewed and extended, with particular reference to cyclonic development at sea level. It is shown that cyclogenesis results from an unbalance, in the layer below the level of non-divergence, of the vorticity advection and the Laplacian of certain thermal quantities. It is suggested that cyclogenesis results not from the release of an infinitesimal perturbation by a dynamically unstable state, but rather from the release of some kind of instability by finite perturbations which can be identified with the wave-shaped motion patterns in the middle and upper troposphere.

1. Introduction

In recent years, Sutcliffe has provided a theoretical treatment of the problem of development which, at least in principle, is quantitative and lends itself to application in forecasting. For details, reference is made to the original papers by Sutcliffe (1939; 1947) and Sutcliffe and Forsdyke (1950).

In his 1947 paper, Sutcliffe tried to combine the effects of temperature advection and vorticity advection into a simple expression. In doing so, he made use of, and substantiated, Dines' model of a quasi-balance in the divergence field, such that boxes of convergence are surmounted by boxes of divergence (and *vice versa*) in such a manner that the mean divergence of any air column, extending from sea level to the top of the atmosphere, is vanishingly small. In numerous ways, Sutcliffe's approach has exerted a stimulating influence on numerical as well as on conventional forecasting.

In an endeavor to develop routine procedures, applicable to present synoptic practices, Sutcliffe assumed the Dines compensation to be effected between the lower and upper troposphere, with a more or less permanent level of non-divergence at about 500 mb. On the basis of various other assumptions and simplifications, Sutcliffe and Forsdyke derived the formula

$$D_0 = \frac{1}{f} V_T \frac{\partial}{\partial s} (2q_0 + q_T + f). \qquad (1)$$

Here D_0 is the isobaric divergence at the 1000-mb level, f the Coriolis parameter, and V_T the thermal wind from 1000 to 500 mb, while s measures length along the streamlines of the thermal wind, and q_0 and q_T are the relative vorticities of the sea-level geostrophic wind and the thermal wind, respectively.

Sutcliffe considered the isobaric divergence as a measure of development, and in the foregoing equation D_0 may be taken as a measure of the rate of development at the 1000-mb level, or sea level. It may be argued that Sutcliffe's definition is not necessarily synonymous with intensification of a moving system. It does, however, provide a measure of the rate of production of vorticity. In the absence of a better definition, we shall adhere to Sutcliffe's nomencalture, realizing that the terms *development* and *divergence* have the same meaning.

Some preliminary experiments with Sutcliffe's technique were made in 1953, and some more extensive experiments, covering a period of about three months, were repeated in early 1954 to determine the usefulness of (1) in routine forecasting. The present article summarizes some notes, written in preparation for these experiments, and includes some experience gained from a cursory examination of a large number of storms and a detailed investigation of a few major storms.[2]

While the results of the experiments with Sutcliffe's technique[3] are difficult to summarize in detail, the following may be said:

1. Equation (1) tends, on the whole, to give undue weight to the sea-level vorticity (q_0); somewhat more appealing results were obtained by omitting q_0 altogether;
2. Almost all major developments examined appeared to occur in qualitative agreement with (1) (with q_0 dropped), but the formula would call for development in many cases where none took place;
3. Equation (1) appears to possess the essence of a necessary criterion, but not that of a sufficient one.

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[2] These investigations were carried out by Miss D. Bradbury, Messrs. P. M. Breistein, M. Estoque, L. L. Means and C. W. Newton, and the writer. Detailed reports will be published at some later time.

[3] The analyses of the Sutcliffe development charts were made by Mr. L. L. Means, of the U. S. Weather Bureau's Chicago Office, who generously assisted the University of Chicago Weather Forecasting Research Center in experimenting with several techniques for predicting cyclogenesis.

2. Diagnostic equation for development

If $\mathbf{V}$, D and Q denote the wind, divergence and scalar absolute vorticity at some pressure level (say, p), respectively, while the same letters with subscript zero refer to a lower level (say, p_0), we may write

$$\mathbf{V} \equiv \mathbf{V}_0 + \mathbf{V}_S, \quad D \equiv D_0 + D_S \quad \text{and} \quad Q \equiv Q_0 + Q_S.$$

Here the subscript S signifies the shear from p_0 to p, and q_S is the relative vorticity of the shear (*i.e.*, $Q = q_0 + f + q_S$).

The vorticity equation for any level may be written

$$dQ/dt = -DQ.$$

Now, since there must be (at least) one value of p at which the divergence vanishes, we have for this particular value $D_S = -D$, and the vorticity equation for this level reduces to

$$\partial q_S/\partial t + \mathbf{V}\cdot\nabla q_S + \mathbf{V}_S\cdot\nabla Q_0 = D_0 Q_0 = -dQ_0/dt. \quad (2)$$

It will be seen that, in the continued absence of differential motion below the level of non-divergence, no development is possible.

If the shear is identified with the thermal wind, and if the lower level is taken at 1000 mb, it follows that no development is possible at this level if the state is barotropic. This result is in agreement with Sutcliffe's simplified formula. On the other hand, (2) shows that baroclinicity is not a sufficient criterion for development.

For diagnostic purposes, it has been customary to replace the actual wind by the geostrophic wind. The advective terms on the left of (2) can then be evaluated from customary synoptic charts. For convenience in writing, we represent the vorticity advection by

$$A_Q = -\mathbf{V}\cdot\nabla q_S - \mathbf{V}_S\cdot\nabla Q_0.$$

The first term on the left of (2) can be obtained from the first law of thermodynamics, in the manner indicated by (3) in the aforementioned paper by Sutcliffe and Forsdyke. Putting

$$\omega = \frac{dp}{dt}, \quad \Gamma_a = \frac{1}{\rho g}\gamma_d, \quad \Gamma = \frac{\partial T}{\partial p}, \quad (3)$$

and representing the temperature advection by

$$A_T = -u\,\partial T/\partial x - v\,\partial T/\partial y,$$

we obtain

$$\frac{\partial q_s}{\partial t} = \frac{R}{f}\nabla^2\left\{\log\left(\frac{p_0}{p}\right) \times \left[\bar{A}_T + \overline{\omega(\Gamma_a - \Gamma)} + \frac{1}{c_p}\overline{\frac{dW}{dt}}\right]\right\}, \quad (4)$$

where dW/dt is the heat energy given to a unit mass per unit time, the bar signifies a mean with respect to intervals of $\log p$, ∇^2 is the horizontal Laplacian operator, and R is the gas constant for dry air. Furthermore, Γ_a and Γ may be interpreted as the adiabatic rate of cooling and the actual lapse rate, respectively, both referred to pressure rather than height.

Equation (2) may now be written as

$$Q_0 D_0 = \frac{R}{f}\nabla^2\left\{\log\left(\frac{p_0}{p}\right) \times \left[\bar{A}_T + \overline{\omega(\Gamma_a - \Gamma)} + \frac{1}{c_p}\overline{\frac{dW}{dt}}\right]\right\} - A_Q. \quad (5)$$

Noting that $Q_0 D_0 = -dQ_0/dt$, one sees that the right-hand side of (5) is a measure of the rate of destruction of vorticity at the 1000-mb level. Since Q_0 is positive, development of cyclonic absolute vorticity at sea level requires that the right-hand side of (5) be negative.

A detailed analysis of the derivations leading to (5) will show that a small term has been omitted, *i.e.*, that which represents the vertical advection of vorticity through the level of non-divergence. At least in the early stages of cyclone development, this should be very small, and its inclusion would hardly throw any additional light on the essential processes associated with development.

The immense complexity of the development processes is apparent from (5). It will be seen that the development at the 1000-mb level comes out as an *unbalance between the vorticity advection and the Laplacian of certain physical properties.* While the vorticity advection is readily obtained from customary synoptic charts, the structure of the thermal contribution is exceedingly complex, and this, perhaps, accounts for the vast variety of developments encountered. As one would expect, the thermal contribution derives from the horizontal advection $\bar{A}_T$, the vertical motion (stability or buoyancy) $\overline{\omega(\Gamma_a - \Gamma)}$, and the non-adiabatic processes, $\overline{dW/dt}$. In addition, it depends upon the height and configuration of the level of non-divergence.

Without endorsing the customary assumption that the level of non-divergence is isobaric (and situated in the middle troposphere), we shall, for the time being, assume an isobaric level and afterwards comment upon modifications due to topography. Since pressure is used as the vertical coordinate, $\log(p_0/p)$ in (5) is to be regarded as a constant as far as the Laplacian operator is concerned.

Restricting our discussion to cyclonic developments of appreciable intensity, we require that D_0 shall be negative and numerically large. Values in the range 10^{-5} to 3×10^{-5} sec^{-1} have been found to be typical during periods of cyclogenesis (see fig. 1).

3. The vorticity advection

With respect to (5), it may be noted that A_Q is positive where the wind is from high to low values of vorticity. Hence, positive vorticity advection in the layer below the level of non-divergence is a favorable condition for cyclone development at sea level. Such vorticity advection is a normal occurrence in advance of troughs in the middle and upper troposphere. While such troughs are always present, cyclogenesis at sea level is a relatively rare event, showing that the thermal contribution must normally be comparable with the effect due to vorticity advection.

On the other hand, experience shows that when low-level cyclogenesis does occur, the locale of development is almost invariably in advance of an upper trough. The vorticity advection is therefore considered to be of primary importance. We shall return to the discussion of this in section 7, in connection with a case of cyclogenesis of rare intensity. A more detailed discussion of the influence of vorticity advection aloft on cyclone development at sea level is given elsewhere in this issue of the JOURNAL (Petterssen *et al.*, 1955).

4. The thermal contribution

Turning next to the thermal contribution, we should emphasize that the various thermal components in (5) are interrelated and cannot be regarded as representing separate physical processes. Nevertheless, we shall comment on the relative importance of these terms to the extent that observational evidence is at hand. As above, we shall assume that the level of non-divergence is isobaric (or nearly so), so that the factor $\log(p_0/p)$ in (5) can be taken outside the Laplacian operator.

Although the non-adiabatic contribution is difficult to evaluate in individual cases, there is ample statistical evidence in support of the view that it is not negligible. The charts of frequencies of cyclogenesis (Petterssen, 1950) show distinct maxima of cyclogenesis and cyclone activity over all non-frozen inland water bodies (surrounded by colder land) in winter. Over these bodies, $\nabla^2\overline{(dW/dt)}<0$, and the non-adiabatic term will contribute to cyclonic development. It may be noted, also, that in summer, when these waters are colder than the surrounding land, $\nabla^2\overline{(dW/dt)}>0$, and the conditions are favorable for anticyclonic development. The above-mentioned frequency charts show that the effect is appreciable.

It is of interest to note that the amount of heating is, in itself, not decisive; what matters is *the configuration of the heating pattern.* Although heating patterns bound to topography may be important locally, they will normally not influence the development of moving systems over lengthy periods.

Loss or gain of heat through radiative processes in the atmosphere may contribute to the development, but it is difficult to visualize that the configuration of these patterns would be such as to render noticeable contributions to the development of extra-tropical cyclones.

Far more important are the non-adiabatic patterns of heating and cooling associated with the horizontal motion (over a thermally non-uniform surface) and the eddy transfer of heat along the vertical. For example, a temperature increase (following the motion) of about 10C a day is not uncommon in polar air streaming southward over northern oceans in winter, while somewhat smaller rates of cooling are observed in tropical air streaming toward higher latitudes. Depending essentially upon the air motion, these patterns are tuned to the system and remain with it while the system moves; the effect is, therefore, a lasting one.

The rate at which the exchange of heat with the underlying surface will affect the layer below the level of non-divergence depends largely upon the lapse rate. On general principles, one would expect this non-adiabatic effect to contribute positively to cyclone development in cold air masses and to have a damping effect in warm masses. (The terms *cold* and *warm*, as used here, refer to Bergeron's classification of air masses.) There is much synoptic evidence to show that the exchange of heat with the underlying surface contributes substantially to cyclone development.

Considering next the buoyancy term [*i.e.*, the term $\nabla^2\overline{\omega(\Gamma a-\Gamma)}$] in (5), we note that no development is possible unless $\bar{\omega}\gtreqless 0$. According to the equation of continuity,

$$D=-\partial\omega/\partial p. \tag{6}$$

Since ω is vanishingly small at sea level and varies monotonically to a numerical maximum at the (first) level of non-divergence, it follows that $\bar{\omega}\gtreqless 0$ if there is any divergence at all. On the same principle, it is readily seen that $\bar{\omega}$ must be negative in cyclonic development ($D_0<0$), and positive in anticyclonic development ($D_0>0$).

Now, in the region of maximum upward motion $\bar{\omega}<0$ and $\nabla^2\bar{\omega}>0$. It follows, then, that the buoyancy term in (5) may be a hinderance or a help, according as the stratification is stable or unstable. In the normal case, when the stratification is stable, the buoyancy constitutes a brake on the development. Disregarding for the moment non-adiabatic influences, we see from (5) that, with stable stratification, development can result only from a favorable combination of the vorticity advection and the Laplacian of the temperature advection.

After vertical circulations have been established and cloud systems formed, it is often found that Γ_a

(now the wet-adiabatic rate of cooling) is smaller than the actual lapse rate Γ. The buoyancy term will then contribute positively to cyclone development. One recognizes here the effect so eloquently advocated by Refsdal (1930). Without saturation, the regions of static instability are relatively small and limited to shallow layers. With saturation, on the other hand, the unstable regions may occupy considerable portions of the vorticity system. In particular, this appears to be true of tropical revolving storms (see Bergeron, 1954).

Forecasting experience in middle latitudes indicates that generally a rich supply of moisture-laden air is necessary to develop a strong extra-tropical cyclone, and this indicates that the buoyancy term may be important. It is, however, difficult to see how the development of extra-tropical cyclones can be *initiated* by saturated over-turning. It appears far more plausible that the buoyancy effect (with saturated instability) comes into play after the development has commenced and large cloud systems have formed as a result of other processes. It is, nevertheless, of interest to note that the static stability affects the development, directly through the buoyancy term and indirectly since it is related to the non-adiabatic processes associated with the earth's surface.

It remains, now, to comment on the temperature-advection term in (5). In the first place, it may be noted that the temperature advection has a positive maximum in advance, and a negative minimum in the rear, of developed storms. Hence, the temperature advection will contribute to cyclone development in advance, and anticyclonic development in the rear, with very little contribution at the center of the storm. It should be noted that the effect of temperature advection is to some extent counteracted by the non-adiabatic influences due to the underlying surface.

The main effect of the temperature advection is to create asymmetry in the production of vorticity, such that cyclonic vorticity is created in advance and destroyed in the rear of the storm, with the result that the storm center obtains a tendency to move in the direction from cold to warm advection. The somewhat vague concepts of *steering* and *blocking* may be discussed on the basis of these considerations, but we shall not do so here.

It is of interest to note that the temperature advection is largely determined by the circulatory motion around cyclones and anticyclones. In the initial stage of cyclone development (*e.g.*, when a frontal wave begins to form), the temperature advection (at least its geostrophic component) is extremely small. It appears, therefore, that the temperature advection is a result of some other developing mechanism. However, when the thermal field has become distorted through circulatory motion, the term $\nabla^2\bar{A}_T$ contributes to further development; in Sutcliffe's terminology, the system has become *self-developing*. This self-development may be further accentuated if $\Gamma > \Gamma_a$ in the cloudy regions.

5. The initial development

Considering now the initial stage of cyclone development, and noting (a) that the term $\nabla^2\bar{A}_T$ is very small initially, (b) that the term $\nabla^2\overline{\omega(\Gamma_a-\Gamma)}$ is opposed to development (before large cloud masses have come in existence), and (c) that the main contribution to the term $\nabla^2\overline{\omega(\Gamma_a-\Gamma)}$ comes from surface activities which depend essentially upon the temperature advection, one's attention is drawn to the possibility that the source of *initial* cyclone development might reside primarily in the vorticity-advection term, and that the remaining terms on the right of (5) might represent processes which become prominent after the development has been initiated.

6. The level of non-divergence

In the foregoing sections, it was assumed that the level of non-divergence was isobaric, so that the factor $\log(p_0/p)$ could be taken outside the Laplacian. Although there is then no question of configuration, the height of the level of non-divergence might be important.

Observations show that the temperature advection is largely limited to the lower troposphere (say, below 600 mb). On the other hand, the term containing $\bar{\omega}$ derives its major contribution from the upper half of the layer below the level of non-divergence. Now, if the level of non-divergence is high (say, at 300 mb), the buoyancy term might dominate the advection term; if the level is low (say, at 700 mb), the reverse may be the case. Observations, to which reference will be made below, show that a high level of non-divergence is typical of cases of non-development and of the initial phase of development, while a relatively low level of non-divergence becomes established as soon as the temperature advection becomes a prominent factor.

It remains now to comment on the configuration of the level of non-divergence. Investigations have revealed that the level of non-divergence varies considerably in space and time and, without further research, it would seem incautious to assume that the level is isobaric. For convenience in writing, we put

$$\Sigma = \overline{A_T} + \overline{\omega(\Gamma_a - \Gamma)} + c_p^{-1}\,\overline{dW/dt}.$$

Since pressure is used as vertical coordinate, p_0 in (5) may be regarded as a constant while p is variable. Equation (5) may now be written as

$$Q_0D_0 = (R/f)[\log(p_0/p)]\nabla^2\Sigma - (R/f)\Sigma(\nabla^2 \log p) - 2(R/f)(\nabla \log p)(\nabla\Sigma) - A_Q, \quad (7)$$

where p is the pressure at the level of non-divergence.

Again, one sees the immense complexity of the development processes, and their dependence upon the height and configuration of the level of non-divergence and the structure of the atmosphere below this level.

Very little is known about the relative magnitudes of the first three terms on the right of (7). In an investigation of a rapidly developing storm (see section 7, below), it was assumed that the non-adiabatic influences could be neglected. ω was computed by the aid of the vorticity equation, and Q_0, D_0 and p were determined. The third term on the right of (7) vanishes where either p or Σ is extreme. For these regions it was found that, although the level of non-divergence was dome-shaped and varied between 400 and 600 mb, the second term on the right of (7) accounted for only about 15 per cent of Q_0D_0. It is impossible to say whether this smallness is typical. In any case, *the usefulness to forecasting of the concept of a level of non-divergence will depend upon how sensitive* (7) *is to variations in the height and in the configuration of the level of non-divergence.* Unless it should prove satisfactory to introduce a standard assumption as to the whereabouts of this level, it would seem difficult to develop routine forecasting procedures based upon Dines' compensation model, although the model would prove useful in theoretical considerations.

In the foregoing sections, use was made of the concept of a level of non-divergence to obtain a relatively transparent expression for the factors that determine development at sea level. It is not suggested, however, that the introduction of this concept is necessary. Since the structure of the atmosphere is overwhelmingly hydrostatic and the motion predominantly geostrophic, development at sea level must be expressible in terms of the conditions of the entire air column. *The introduction of a level of non-divergence only serves to eliminate the need for knowledge of the conditions above this level;* instead, it introduces the need for knowledge of the whereabouts of the level itself.

7. A case of intense development

Although it is not the purpose of this article to present observational data, fig. 1 is included to illustrate a case of intense development.

On 23 November 1952, a feeble frontal wave was present over northeastern Texas, while an intense trough in the upper troposphere was situated over the Pacific Coast and the Rocky Mountains. No appreciable development occurred until the morning of the 25th, when the upper trough (with strong vorticity advection in advance of it) approached the frontal wave. The development at sea level set in suddenly, just after 0600 GCT. The resulting storm moved northeastward, on an almost straight path (see fig. 2), and crossed James Bay on the 27th. The whole development was accomplished in about 24 hr, during which time the relative vorticity at the center increased by a factor of more than four, while vast amounts of vorticity were created and spread over large areas surrounding the center. The storm certainly represented a major meteorological event.

The lowest part of fig. 1 shows the absolute vorticity over the sea-level storm center while it moved. Prior to the development, the "vorticity axis" slanted strongly toward the west; it gradually became steeper, and approached an almost vertical position at the end of the development (see fig. 2).

The top part of the diagram in fig. 1 shows the vertical velocity (*i.e.*, dp/dt) over the storm center, and in the middle is shown the isobaric divergence. The heavy lines, in all three parts of the figure, indicate zero divergence.

The magnitude of the vorticity advection aloft is

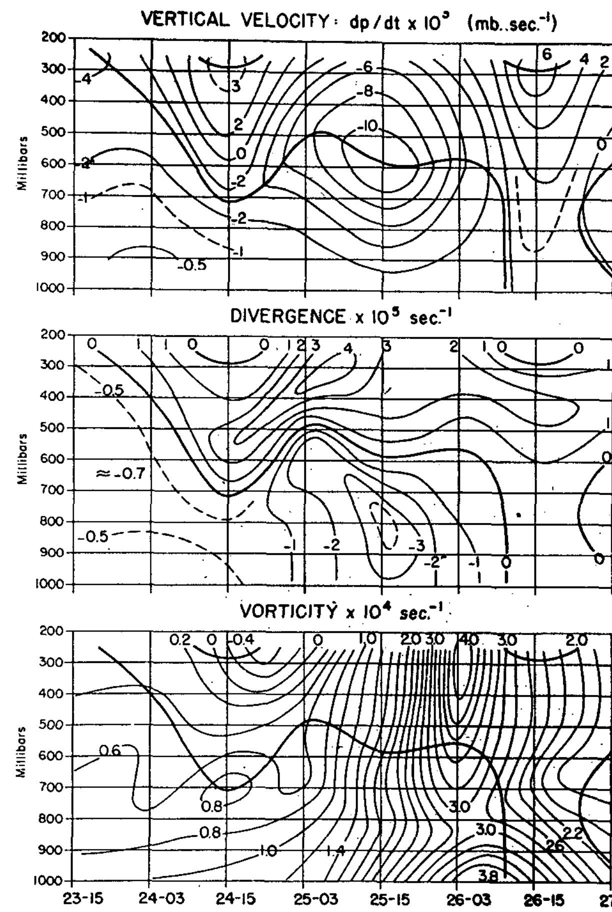


FIG. 1. Distribution of absolute vorticity (Q), divergence (D), and vertical velocity (dp/dt) in central column of frontal wave which developed into major storm, 23–27 November 1952. Path is shown in fig. 2.

readily visualized from the lowest diagram of fig. 1 and from fig. 2. It will be seen that the vorticity in the upper divergence area over the storm center increased very rapidly. This vorticity was created in the convergence area to the west of the approaching upper trough, and advected (with some loss due to divergence) into the area over the center.[4]

It will be seen from fig. 1 that, on the 23rd, the level of non-divergence was in the upper troposphere, and the Dines compensation was then effected between the troposphere and the stratosphere. The level descended and varied around about 600 mb on the 24th, while a second level became discernible above 300 mb. It is immediately apparent that, on the 24th, the Dines compensation was effected *mainly* between the middle and upper troposphere on the one hand, and the stratosphere on the other. In other words, the *main* level of non-divergence was still above the 300-mb level.

In the early hours on the 25th, there was a sudden change in the compensation, such that, temporarily, an almost complete balance existed between the lower and upper troposphere. The level remained between 600 and 500 mb until the morning of the 26th, when it descended abruptly to sea level. The development was then terminated; a new level of non-divergence appeared at great heights, and compensation between the troposphere and the stratosphere was re-established.

The events shown in fig. 1 may be referred to four periods, as shown in table 1. Since the mean divergence of the entire column must be vanishingly small, the contribution at levels higher than 300 mb may be computed from the values below this level. It will be seen that, except during the period of rapid development, the main compensation was between the troposphere and the stratosphere, and that the amounts of divergence in the stratosphere were appreciable. However, if the mean for the entire period is considered, one finds the main level of non-divergence at about 600 mb, and no significant contribution from the levels above 300 mb.

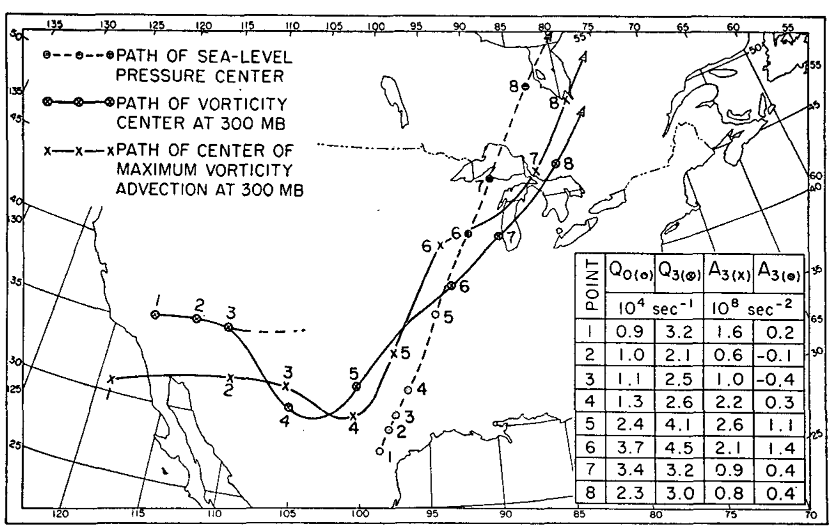
[4] A more detailed description of this storm is given elsewhere in this issue of the JOURNAL (Petterssen *et al.*, 1955).

TABLE 1. Divergence (10^{-5} sec^{-1}) in column over cyclone center, 23–27 November 1952. A: initial quasi-stationary conditions; B: forward march of upper trough; C: period of intense development; D: period of slow dissipation.

Pressure (mb)	*A*	*B*	*C*	*D*	Mean (*A* to *D*)
1000–900	−0.3	−0.5	−2.2	0.0	−0.75
900–800	−0.5	−0.6	−2.3	0.0	−0.85
800–700	−0.6	−0.5	−2.2	0.1	−0.80
700–600	−0.7	0.3	−1.5	0.3	−0.40
600–500	−0.6	1.6	−0.1	0.9	0.45
500–400	−0.5	1.8	1.7	1.6	1.15
400–300	−0.2	1.0	2.8	1.4	1.25
300–0	3.4	−3.1	3.8	−4.3	−0.05
Main level of non-divergence	Above 300	Above 300	At 550	Above 300	

It was mentioned in section 3 that almost all major developments at sea level occur in advance of advancing troughs in the upper troposphere. The storm discussed briefly above is typical in this respect.

In fig. 2 are shown the positions of the sea-level center at 12-hr intervals from 1500 GCT 23 November, to 0300 GCT 27 November, and the corresponding positions of the center of maximum vorticity at the 300-mb level (associated with the upper trough), and, also, the positions of the center of maximum vorticity advection at the 300-mb level. The inset table shows (column-wise, from left to right) the absolute vorticity at the sea-level pressure center, the absolute vorticity at the vorticity center at the 300-mb level, the vorticity advection [$A_{3(x)}$] at the center of maximum vorticity advection at 300 mb, and the vorticity advection at the 300-mb level over



POINT	$Q_{0(\circ)}$	$Q_{3(\otimes)}$	$A_{3(x)}$	$A_{3(\otimes)}$
	10^4 sec^{-1}		10^8 sec^{-2}	
1	0.9	3.2	1.6	0.2
2	1.0	2.1	0.6	-0.1
3	1.1	2.5	1.0	-0.4
4	1.3	2.6	2.2	0.3
5	2.4	4.1	2.6	1.1
6	3.7	4.5	2.1	1.4
7	3.4	3.2	0.9	0.4
8	2.3	3.0	0.8	0.4

FIG. 2. Path of storm referred to in fig. 1, and certain vorticity and vorticity-advection characteristics.

the sea-level center. The suffixed symbols in the column headings are those used to identify the paths.

It will be seen that, during the first 36-hr period (points 1 to 4), the upper trough, with its maximum of vorticity and vorticity advection, approached the sea-level system. During this period, there was no significant development at sea level; there was no significant vorticity advection over the sea-level center, and the advection further to the west was apparently compensated for by cold advection to the south and east of a high situated over the northern Rockies. During the following 24-hr period, a rapid development of vorticity at sea level took place; the vorticity advection over the sea-level center was then large, and there was no significant temperature advection in the central column. A maximum of positive temperature advection was present to the northeast of the sea-level center; this added to the effect of the vorticity advection and probably accounted for accelerated movement of the center, as indicated by positions 3 to 5 in fig. 2.

The maximum intensity of the sea-level center was reached shortly after point 6. At about this time, the path of the vorticity advection center aloft crossed the path of the sea-level center; the vorticity advection over the center decreased suddenly, and cold advection spread into the central column while the system occluded.

In this particular storm, there can be little doubt that the initial and sudden release of the development at sea-level was due mainly to the vorticity advection, while the thermal advection became effective shortly after the development had commenced.

8. The importance of fronts

It is well known that almost all cyclogenesis in middle and high latitudes occurs in connection with fronts, and one may ask whether the existence of a frontal surface is an essential requirement. It was shown in section 2 that baroclinicity (or, more precisely, shearing motion) is essential, and this emphasizes the importance of fronts. Furthermore, the temperature varies sharply across the frontal region, and as soon as a circulatory motion is created (*e.g.*, through unbalanced vorticity advection, see section 5), a self-developing situation is established through the resulting temperature advection.

It is of interest to note that, since the upper troughs are cold, the forward march of such a trough relative to the sea-level frontal system (see section 7) essentially means a steepening and intensification of the frontal surface or, in other words, a frontogenesis in depth. It appears, therefore, that both the existence of a frontal surface and an intensification of the frontal region are important factors. While frontogenesis in the free atmosphere is difficult to diagnose, the forward march of an upper trough can readily be diagnosed and prognosticated with some success. With use of the vorticity advection ahead of advancing upper troughs and the presence of a frontal zone as sole criteria for development, an experiment with forecasting of cyclone development has been conducted. The results are described elsewhere (Petterssen *et al*,. 1955).

9. Infinitesimal or finite perturbations

In connection with the storm referred to in section 7, it is of interest to note that the upper trough (with positive vorticity advection in advance of it), which appeared to release the sudden and intense development, was part of a wave whose length and amplitude were about 5000 and 3000 km, respectively. Regarded as a perturbation superimposed upon the zonal curent, it was one of formidable dimensions. According to the classical perturbation theory, cyclone development should result from the release of an infinitesimal perturbation by a dynamically unstable state. In the case referred to in section 7, and in other cases examined, it appears that the unbalance (or instability, if it may so be called) was released by finite perturbations. The results of the experiment referred to above indicate clearly that it is often possible to diagnose and prognosticate likely regions of cyclogenesis from finite factors.

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REFERENCES

Bergeron, T., 1954: The problem of tropical hurricanes. *Quart. J. r. meteor. Soc.*, **80**, 131–164.

Petterssen, S., 1950: Some aspects of the general circulation of the atmosphere. *Centenary Proc. roy. meteor. Soc.*, 120–155.

——, G. E. Dunn, and L. L. Means, 1955: Report on an experiment in forecasting of cyclone development. *J. Meteor.*, **12**, 58–67.

Refsdal, A., 1930: Der feuchtlabile niederschlag. *Geofys. Publ.*, **5**, No. 12, 71 pp.

Sutcliffe, R. C., 1939: Cyclonic and anticyclonic development. *Quart. J. r. meteor. Soc.*, **65**, 518–524.

——, 1947: A contribution to the problem of development. *Quart. J. r. meteor. Soc.*, **73**, 370–383.

——, and A. G. Forsdyke, 1950: The theory and use of upper air thickness patterns in forecasting. *Quart. J. r. meteor. Soc.*, **76**, 189–217.