551.515.1: 551.8(54)

On the Formation of Monsoon Depressions in the Bay of Bengal

P. KOTESWARAM and C. A. GEORGE

Meteorological Office, Calcutta

(*Received 14 March 1957*)

ABSTRACT. The role of the upper tropospheric perturbations in the formation of monsoon depressions in the Bay of Bengal has been examined. From a study of these flow patterns, it is found that cyclonic development at sea level occurs when and where an area of positive vorticity advection in the upper troposphere becomes superimposed upon a pre-existing trough at sea level. Other factors affecting cyclogenesis at sea level during the southwest monsoon period are also pointed out.

1. Introduction

Tropical depressions forming in the Bay of Bengal and occasionally in the Arabian Sea during the southwest monsoon season (June to September) are well known for more than half a century. These monsoon depressions are low pressure areas with a pressure departure of nearly 2–6 mb at the centre with cyclonic winds not exceeding 30 knots in speed. If the wind speed is between 30 and 40 knots and the pressure departure is 6–10 mb they are classified as cyclones. They do not become severe cyclones reaching hurricane intensity. They are fairly large in extent compared to tropical cyclones occurring in the pre-monsoon and post-monsoon seasons and sometimes affect an area as large as 10^5 square miles. Eliot (1900) who mentions them in his *Handbook of Cyclonic Storms*, considered that the weather in monsoon depressions is determined by the interplay between two air streams, one a dry one and another a deep moist one (the SW monsoon). Climatological details of regions of their formation, frequency of occurrence and tracks followed by them are available in the periodical publications of the India Meteorological Department. Their role in the distribution of rainfall over the sub-continent in the SW monsoon season has been recognised and numerous attempts have been made by Indian meteorologists to study their structure and the conditions of their formation.

1·1. *Air mass concepts* — The earliest attempt to present a hypothesis for the formation of these depressions was made by Roy and Roy (1930). They and the subsequent workers were strongly influenced by the Norwegian models for extra-tropical depressions postulated at the beginning of the century. Air masses were identified and discontinuities or fronts looked for. Roy and Roy (1930) opined that three air masses—fresh monsoon air, dry continental air from the northwest and monsoon air deflected and dessicated by the hills of northern India—took part in the formation of depressions at the head of the Bay of Bengal. Ramanathan and Ramakrishnan (1932) found that the monsoon air that streams across the Peninsula accelerates prior to the formation of a north Bay depression and considered that depressions form at the discontinuity between the fresh monsoon air and the turned or old monsoon air to the north. The Burmese mountains to the east help in rolling up the monsoon current into a vortex. The effect of these mountains is, however, obscure, since depressions are known to form even in the western half of the Bay far away from the mountain ranges. Malurkar (1950) gave a Pacific Ocean origin to the easterly current to the north of the depression and called it *Tr* (far east transitional air). He considered that this air mass as well as *Em* (SW monsoon) and *Tc* (continental air from the northwest) were essential for the formation of the depressions.

All these hypotheses were based upon surface observations and some pibal data, which were mainly confined to about 10,000 feet a.s.l. due to the extensive cloud cover during the monsoon period. Occasional sounding balloon flights were, however, made for studying particular disturbances and the existence of fronts confirmed from available temperature records (Sur 1932). After routine radiosonde observations became available since 1944, attempts were made to identify fronts and air mass discontinuities with the aid of upper air temperature data. Mean characteristics of the different air masses in the Indian area were studied by Roy (1946), who identified the easterlies over north India during the monsoon season as turned monsoon air (*EmT*). Due to the poor temperature contrasts between the air masses supposed to take part in the depressions and the absence of discontinuities in the radiosonde ascents characteristic of extra-tropical fronts, doubts were sometimes raised regarding the existence of these fronts (Pramanik and Rao 1948, Mull and Rao 1949). Desai (1950) made a detailed study of a Bay depression in 1947 with the aid of radiosonde data and came to the conclusion that it formed on the discontinuity between the monsoon air mass and a mixture of subsided tropical maritime air from the far east and the turned monsoon air. Desai and Koteswaram (1951) confirmed the earlier frontal models in the case of a monsoon depression of July 1945.

1·2. *The perturbation approach*—During and after World War II when flying activity over some parts of the tropics was extensive, detailed observations were collected over the tropical Pacific and the Carribean areas. These data did not support frontal concepts. Riehl (1945) made a perturbation approach for the study of tropical weather. He formulated models of waves in the easterlies which, under favourable circumstances, intensified into typhoons or hurricanes (Riehl 1948). Palmer (1951) made a searching enquiry into the various methods of tropical analysis and stressed the untenability of frontal concepts in the tropics. As pointed out by Riehl (1954 b) the fundamental difference between the deepening cyclones in the tropical and extra-tropical areas is that all air entering the circulation rises in the case of the tropical disturbances, while only tropical air rises and polar air sinks in extra-tropical circulations.

1·3. *Dynamical models*—With the application of dynamical concepts in the field of synoptic meteorology during recent years, much progress has been made in the understanding and forecasting of atmospheric disturbances. Even in the extra-tropical latitudes, the simple Bjerknes frontal models have been found to be inadequate either for indicating development or fully explaining the rainfall patterns associated with depressions. An old concept of mass compensation between the upper and lower layers of the troposphere as suggested by Dines (1912) in order to account for the presence of divergence over developing cyclones and convergence over anticyclones has been revived in recent years. Bjerknes (1937) drew attention to the role of unstable growth of an upper wave trough in inducing cyclogenesis at sea level (see also Bjerknes 1951). Sutcliffe (1947) obtained an expression for the rate of development in terms of vertical velocity and divergence. He took divergence as a measure of development and showed that a part of the effect of convergence would intensify the vorticity system and the other part would advect it relative to the moving system. Sutcliffe and Forsdyke (1951) obtained the surface divergence as equal to but opposite to the thermal divergence between the sea level and the layer of non-divergence (assumed to be 500-mb level). Petterssen (1955) amplified Sutcliffe's ideas and obtained a complete equation for development as follows—

$$\dot{Q}_0 = -Q_0 D_0 = A_Q + \mathbf{V}_0 \cdot \nabla Q_0 - \frac{R}{f} \nabla^2 \left(\frac{g}{R} A_T + S + H \right) \quad (1)$$

where Q_0 is the vertical component of absolute vorticity at sea level, D_0 is the divergence at sea level, A_Q and A_T indicate the advection of vorticity and thickness, S and H are terms for stability and heat and other terms have their conventional meanings.

The complexity of the development process is evident. Since the sea level advection $\mathbf{V}_0 . \nabla Q_0$ is generally found to be small, development at sea level arises out of an "imbalance between the vorticity advection at the level of non-divergence and the Laplacian of the thermal components A_T, S and H" (Petterssen 1955). Petterssen arrived at a simple hypothesis that "cyclonic development at sea level occurs when and where an area of positive vorticity advection becomes superimposed upon a frontal zone at sea level" (Petterssen, Dunn and Means 1955).

Riehl (1954a) found that when a pronounced jet stream became superimposed on a strong frontal zone or some perturbation in the lower troposphere, intense cyclogenesis was caused. He studied the effect of jet fingers or jet maxima and found that deepening of the lower tropospheric low was associated with a jet maximum and occurs on its left forward sector. The greatest deepening occurred when the low reached a position east of the long wave trough in the westerly jet axis.

It would be clear from the foregoing, that models mainly based on low tropospheric features like those developed in the Indian area so far, are necessarily incomplete and any hypothesis for tropical disturbances should take into account the contributions of both the lower and the upper troposphere for sea level developments. With this end in view, a detailed analysis of upper and lower tropospheric perturbations leading to the formation of depressions in the southwest monsoon season was attempted and the results are given in the following paragraphs.

2. The structure of the upper troposphere over India during the southwest monsoon season

The importance of the study of upper tropospheric conditions for the understanding of low tropospheric disturbances has been sufficiently stressed in the preceding section. Such a study is not possible unless there is an adequate coverage of radiosonde and rawin data at least upto the tropopause level. Radiosonde data are available in India for more than a decade; but it is not possible to make reliable conclusions based upon daily radiosonde observations alone, since the magnitude of errors involved in observing the various elements are comparable with the gradients observed on constant pressure maps during the summer season over the tropics. Rawin data could provide more dependable aids. Rawin stations have been started at various locations in the country since 1954 and upto date there are 10 such stations over India—quite a meagre network still.

With the aid of available rawin data for India and the network of rawin stations elsewhere in the northern hemisphere, Koteswaram (1956) made a study of the upper tropospheric flow over the tropics during summer. As is well-known, this flow pattern is mostly easterly in summer. While over the Pacific and Atlantic Oceans it breaks up into vortices (Riehl 1948), over south Asia it is more organised and consists of a huge anticyclone based over the Tibetan plateau with an easterly jet stream at its southern rim starting from the China Sea and extending into North Africa. The jet which runs roughly along Lat. 15°N at about 150-mb level accelerates from China Sea to South India and decelerates thereafter. Wind speeds of 100 kts or more are reached in its core. Since the right entrance and the left exit of the jet are associated with upper vorticity advection and consequent vertical ascent, he finds a relation between the activity of the lower monsoon and the location of the jet. The upper current thus provides an important clue to the development of monsoon weather in the lower troposphere.

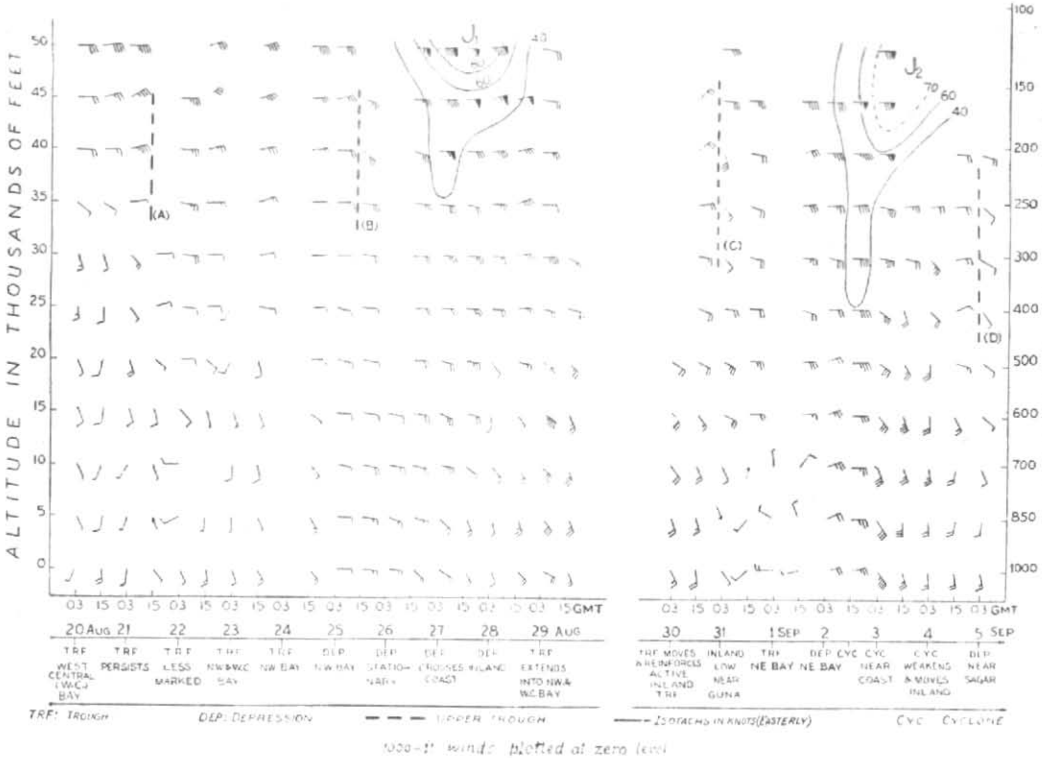


Fig. 1(a). Time-section for Calcutta (20—29 August 1955)

Fig. 1(b). Time-section for Calcutta (30 August—5 September 1955)

2·1. *Upper Easterly Waves*—During the course of the above study, it was noticed that the upper jet stream was of a pulsating character and shifts its position both laterally and longitudinally. The breaking up of the current into a number of jet streaks or jet fingers as in the case of the westerly jet stream could be inferred but could not be substantiated due to lack of adequate data. Apart from these perturbations, evidence was available for wave-like disturbances moving from east to west in the upper troposphere between 500 and 300 mb. These waves in the upper easterlies are, however, generally damped with height and at 200 mb and above the flow is mainly straight on most days, with fluctuations reflected in the wind speed in the upper jet streams. When there is no strong current in the highest levels of the troposphere, the easterly waves can be seen extending upto the tropopause.

Figs. 1(a) and 1(b) give the vertical time-sections of Calcutta for the period 20 August to 5 September 1955, where the passage of these upper easterly waves (marked by heavy broken lines) and wind maxima (indicated as J_1 and J_2) have been illustrated. It is also observed that the heights of isobaric surfaces fall in the upper levels before the approach of the wave trough and rise in its rear. Similar wave passages have been noticed in the time-sections of other stations in the Indian monsoon area like Madras and Bombay as well as at Singapore and Aden. From an examination of hemispherical charts during 1955, it was found that most of these waves had their origin in the southwest Pacific and some of them could be traced from even Long. 140°E.

3. Influence of upper tropospheric perturbations on the formation of monsoon depressions in the north Bay

With a view to observe the effects, if any, of these upper easterly waves as well as the fluctuation in the jet stream on the formation of monsoon depressions in the north Bay, a detailed analysis of available rawin data for Calcutta from 1954 to 1956 and Madras from 1955 to 1956 was undertaken. Continuous height-time sections of winds were constructed for the entire monsoon period June to September each year and the passages of the wave troughs as well as the wind speed maxima in the upper air over Calcutta were noted. The formation of unsettled conditions, depressions or cyclones in the north Bay was also indicated on the relevant dates. The passage of the upper easterly waves over Calcutta was confined in most cases with its passage over Madras a day or so later.

3.1. *Upper Easterly Waves and Monsoon Depressions*—During the three monsoon seasons 1954–56, 37 easterly wave troughs passed over Calcutta; and on an average, they were found over Calcutta with a frequency of one in about six days during the height of the monsoon.

The following sequences were observed in association with these upper wave troughs over Calcutta —

(*i*) The passage of an upper easterly wave over Calcutta was preceded always by the formation of unsettled conditions or the extension of the seasonal trough over the north Bay.

(*ii*) With a pre-existing trough or an extension of the normal monsoon trough in the north Bay, the approach of a fresh easterly wave resulted in the formation of a monsoon depression. The depression was formed in the northeast Bay prior to the appearance of the wave trough over Calcutta and in the northwest Bay almost simultaneously with its passage over Calcutta.

(*iii*) The upper trough generally overtakes the lower depression and when the trough has passed over the centre, the depression weakens.

3·2. *Easterly Jet Stream and Monsoon Depressions*—As already mentioned, the easterly jet stream is located at the 150-100 mb level during the southwest monsoon season and generally runs near about the latitude of Madras. However, strong winds are also observed occasionally as far north as Calcutta. In the time-section presented in Fig. 1(a) winds exceeding 50 knots occur at Calcutta above 200 mb during the period 26 to 28 August 1955 soon after the passage of the wave trough on 25th. A similar strengthening of the winds may be noticed in Fig. 1(b) where there is no corresponding wave trough movement in the lower levels but a wind maximum passed over Calcutta on 2 and 3 September. An examination of the time-sections of Calcutta during the three years 1954–56 revealed that there were 11 occasions during the monsoon season when the upper winds over Calcutta at the jet level (150-100 mb) exceeded 50 knots. The surges are not periodic and sometimes may last for even 15 days over the station (*e.g.*, June–July 1955). These strong winds are probably due to the passage of a jet streak over Calcutta but it is not possible to locate the jet accurately in the absence of adequate observations. As indicated earlier, these occasional surges are most likely caused by the passage of jetlets or subsidiary wind maxima over Calcutta. Such breaking up of the main current into jet fingers or jetlets is well known in the westerly jet stream and is probably a property of the easterly current also.

With the passage of these wind maxima or jetlets over Calcutta, sequences similar to those pointed out in the previous section have been noticed. With the approach of the wind maximum, pre-existing troughs at sea level in the north Bay intensify into depressions and circulations weaken after the wind maximum has overtaken the low level vortex (*see* Fig. 1 b). Development associated with these wind maxima was, however, found to be more rapid than in the case of the upper easterly waves.

Table 1 gives the observed relation between the high level Easterly Waves (E.W.) and Easterly Jet maxima (E.J.) over Calcutta

TABLE 1

Number of occasions when monsoon depressions were formed in the North Bay in association with the passage of easterly waves and jet maxima over Calcutta during the monsoon period of 1954-56

Month	No. of cases of passage of E.W.	No. of cases of passage of E. J.	No. of cases with sea level trough	No. of cases without sea level trough	No. of depressions formed by E.W.	No. of depressions formed by E. J.
(1)	(2)	(3)	(4)	(5)	(6)	(7)
June	4	4	6	2	2	4
July	9	5	5	9	3	2
August	11	1	5	7	4	1
September	13	1	9	5	8	1

and the formation of monsoon depressions in the north Bay during the three years 1954–56. The numbers in column 4 indicate the number of occasions when the trough at sea level had extended into the north Bay before the passage of the upper perturbation. Cases when the monsoon trough had extended into the north Bay when there were no upper perturbations have not been included in the table, since on none of these occasions the monsoon trough had intensified into a depression.

The correspondence between the numbers in col. 4 and the totals of cols. 6 and 7 provides conclusive evidence regarding the mechanism involved in the formation of the depressions. It is clear that *with the approach of an easterly wave or wind maximum in the upper troposphere, a pre-existing low at sea level over the north Bay of Bengal would invariably intensify into a depression.* When no such pre-existing low is present, *i.e.*, when the monsoon trough had not already extended into the north Bay, the passage of an upper trough or wind maximum did not result in a depression at sea level; at best, it caused unsettled conditions or heavy rains. It may be mentioned here that Pisharoty and Kulkarni (1956) have noticed the occurrence of heavy rain over western India in association with similar passages of upper westerly waves over north India in the early monsoon season.

4. Case studies of some monsoon depressions

The conclusion in the previous paragraph was tested by examining the synoptic conditions which preceded the formation of 8 monsoon depressions in the north Bay of Bengal. Charts were drawn for the 1000, 700, 500, 300 and 200 mb-levels. Above 300 mb, streamline analysis was preferred to contour analysis due to the unreliability of radiosonde data. The formation of two typical depressions is described in the succeeding paragraphs.

4·1. *Depression of 25–28 August 1955*—The monsoon trough extended into the north Bay on 23 August 1955 and intensified into a depression on the 25th with its centre at 0300 GMT near Lat. 18°N, Long. 87°E. It intensified into a deep depression during the next 24 hours and crossed coast near Gopalpur (Lat. 19°N, Long. 85°E approx.) on the evening of the 27th.

The above development is indicated in Fig. 1(a), which illustrates the easterly wave from 250 to 150 mb with a wind maximum aloft following the wave. The relevant 200-mb flow patterns and the sea level isobars for the Indian area are shown in the map sequences in Figs. 2(a) to 2(f).

On the morning of 22nd, an easterly wave trough marked (*a*) in Fig. 2(a) at the 200-mb level lay off the coast of China roughly along Long.115°E. This trough moved inland during the next 24 hours and was located on 23rd roughly along Long. 105°E as could be judged from the winds at Saigon (100°, 46 knots) and Singapore (070°, 53 knots). The wind at Hongkong veered from 050°, 47 knots on 22nd morning to 130°, 21 knots on the morning of 23rd. By 0300 GMT of 24th, the upper trough had probably moved further westwards. Obervations from SE Asia were absent; but from the previous movement of the wave it could be inferred to lie approximately along Long. 95°E on the

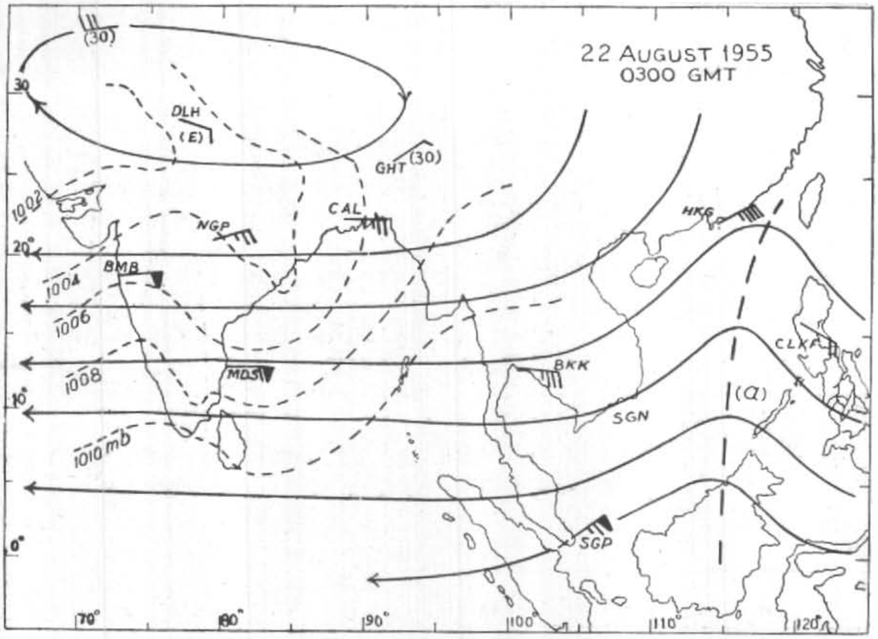


Fig. 2(a)

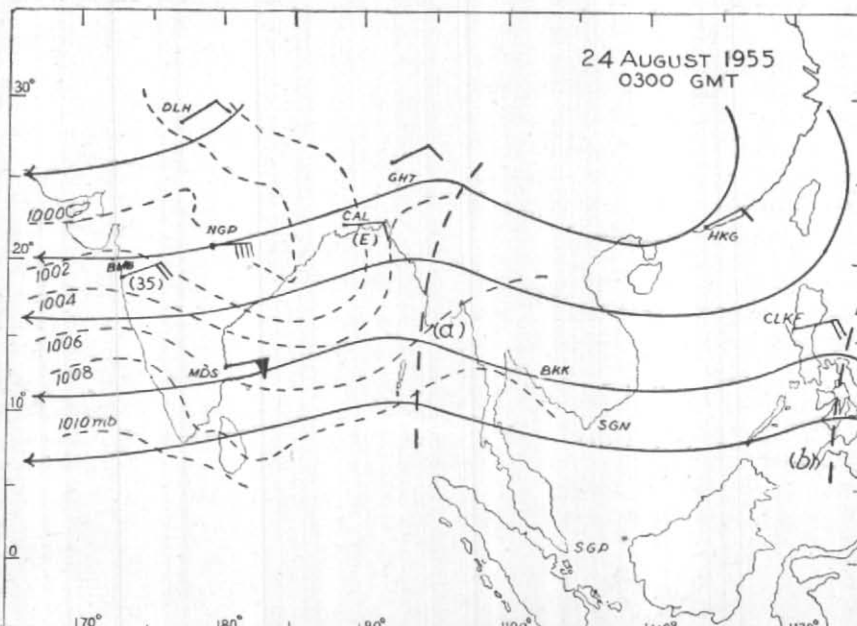


Fig. 2(b)

Figs. 2(a) and 2(b). Flow patterns at the 200-mb level based on the 0300 GMT rawin data

—— Stream lines, ------- Sea level isobars at 0300 GMT, — — — — Wave troughs

E—Evening winds. Figures within brackets indicate the level of the wind in thousands of feet

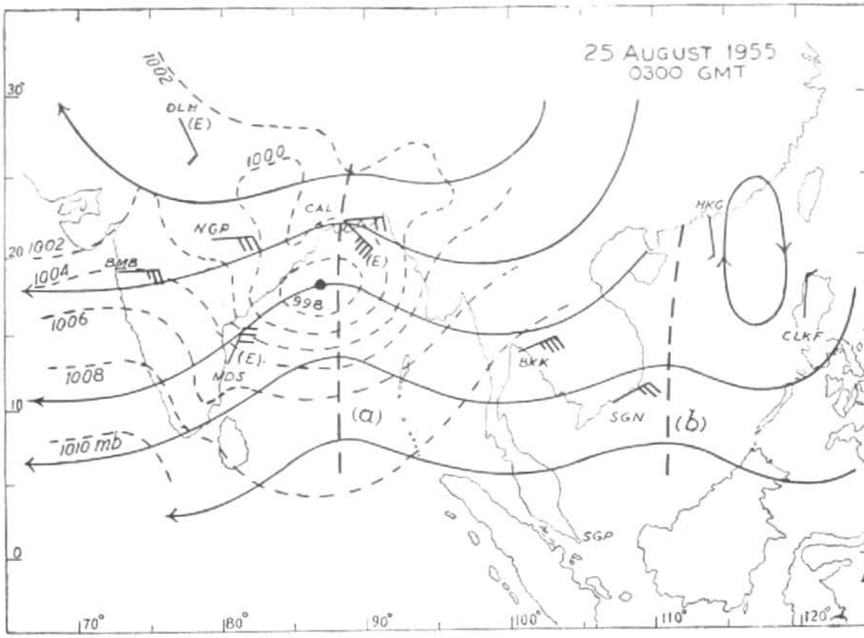


Fig. 2(c)

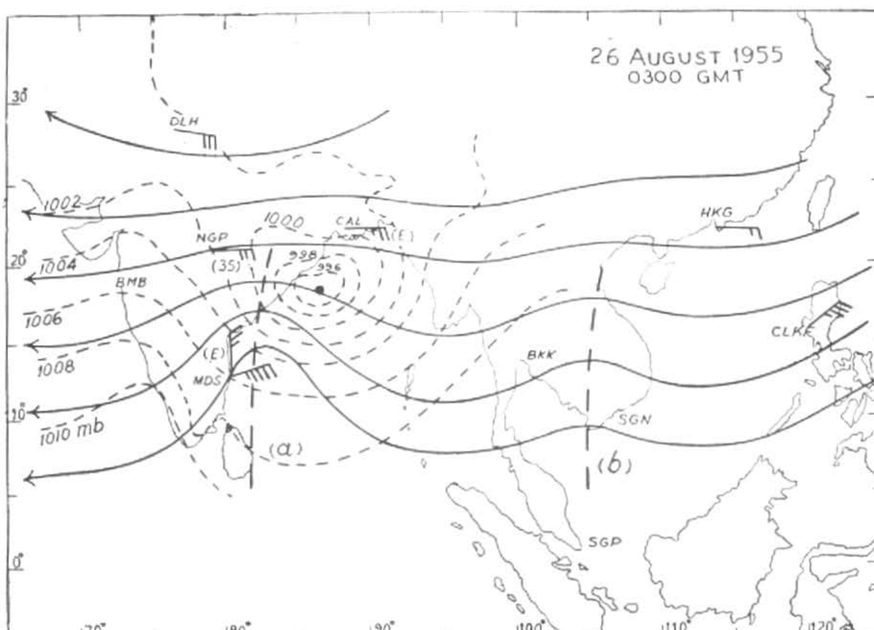


Fig. 2(d)

Figs. 2(c) and 2(d). Flow patterns at the 200-mb level based on the 0300 GMT rawin data

—— Stream lines, -------- Sea level isobars at 0300 GMT, — — — — Wave troughs

E—Evening winds. Figures within brackets indicate the level of the wind in thousands of feet

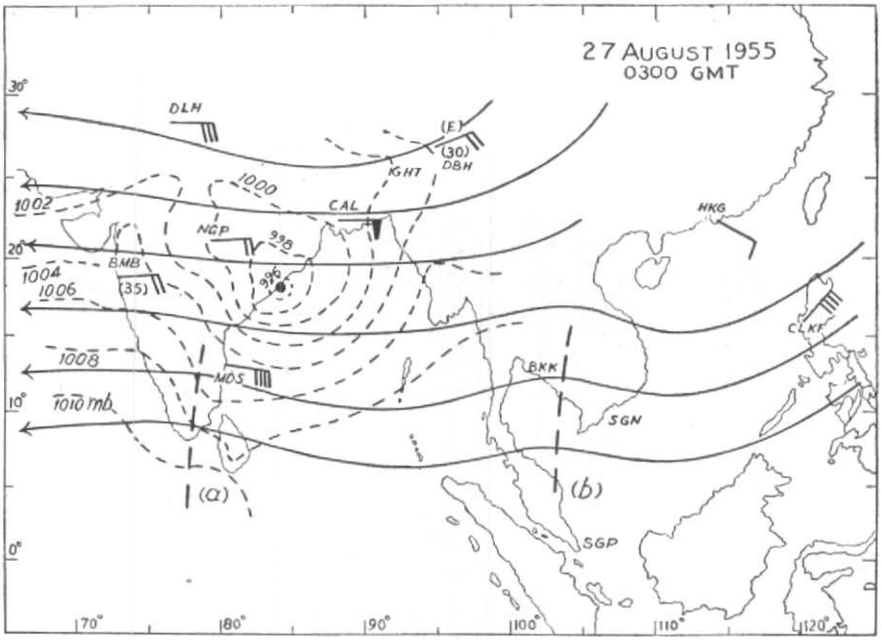


Fig. 2(e)

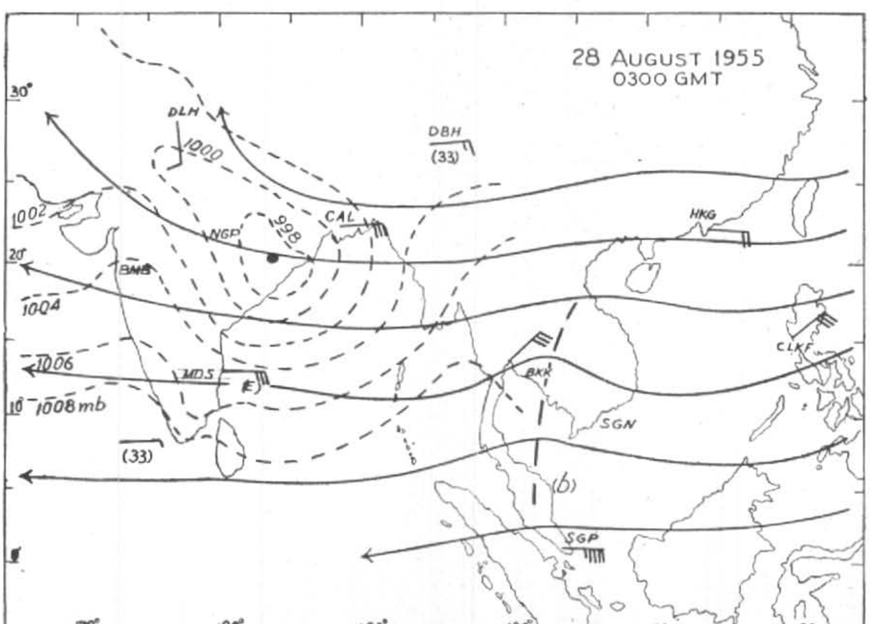


Fig. 2(f)

Figs. 2(e) and 2(f). Flow patterns at the 200-mb level based on the 0300 GMT rawin data
—— Stream lines, ------ Sea level isobars at 0300 GMT, — — — — Wave troughs
E—Evening winds. Figures within brackets indicate the level of the wind in thousands of feet

morning of 24th (Fig. 2b). The sea level trough can be seen extending well into the northwest Bay. On the 25th, the 200 mb trough moved into the north Bay and Calcutta winds veered from 080°, 19 knots to 130°, 36 knots from the morning to evening of this day. From 24th to 25th, the low level trough came under the influence of the forward sector of the advancing upper easterly wave and by the morning of 25th, it intensified into a depression with central region near Lat. 18°N, Long. 87°E. Its position with respect to the upper wave may be seen in Fig. 2(c). It further intensified into a deep depression with the passage of the upper wave trough which overtook the lower depression during the next 24 hours. Fig. 2(d) illustrates the upper wave trough and the low level depression on the morning of 26th. The wind reported by Madras on the morning of 26th (070°, 44 knots) has been rejected as doubtful as compared to the observations at 1500 GMT of 25th and 26th (030°, 28 knots and 010°, 26 knots respectively). Hence the upper trough may be assumed to be located to the east of Madras on the morning of 26th (and even on the evening) near about Long. 82°E. The movement of the upper trough apparently slowed down. With the passage of the upper trough westwards, winds over Calcutta increased from 090°, 25 knots at 0300 GMT of 25th to 090°, 56 knots at 1500 GMT of 26th at the 100-mb level. This increase was maintained during the next two days, after which they weakened (*see* Fig. 1 a). The sharp increase of winds at the jet stream level over Calcutta would indicate the arrival of an upper jet maximum on the 26th. It is quite likely that the cyclogenesis in the northwest Bay was influenced not only by the wave trough but also by the approaching jet maximum between 150 and 100-mb levels.

Though the depression came under the influence of the rear sector of the upper wave on the 26th, it did not change appreciably either in intensity or in its position, apparently due to the continuing effect of the approaching jet maximum above 150 mb. The 200-mb trough crossed Madras meridian between 1500 GMT of 26th and 0300 GMT of 27th, when Madras winds changed from 010°, 26 knots to 110°, 38 knots. By 27th, the upper wave probably got damped and straight easterly flow was observed at the 200-mb level (Fig. 2 e). The depression which had moved westnorthwest, was on the morning of 27th just crossing the coast near Gopalpur. It was now under the influence of the jet above 150 mb. Calcutta recorded 090°, 88 knots at 100 mb on the morning of 27th and 080°, 60 knots on the same evening. On 28th morning it again reported 080°, 86 knots. The maximum had passed over Calcutta between 27th and 28th. The depression had crossed coast on the evening of 27th and probably came under the influence of the rear of the wind maximum. Calcutta wind fell to 100°, 20 knots at 100 mb level on the morning of 29th. The depression moved inland by 28th morning (Fig. 2 f) and weakened into a low pressure area between 28th and 29th.

4·2. *The cyclone of 2 September 1955*—The period between 20 August and 5 September 1955 was very favourable for cyclogenesis. In Figs. 1(a) and 1(b), the passage of 4 waves and two jet maxima has been indicated. The effect of the trough marked B as well as the jet J_1 has been discussed in the previous section. On the arrival of the trough C, conditions again became unsettled (Fig. 1b) in the northwest and west central Bay on the 29th and a closed low could be located over the area. This did not intensify but rapidly passed inland, where it merged with the remnant of the previous depression and formed a fresh low near Guna (Lat. 24½°N, Long. 77½°E) by the morning of 31st. This low moved away northwestwards.

With the approach of the second jet maximum J_2 of the period (*see* Fig. 1 b), the monsoon trough again extended into the northeast Bay by the morning of 1 September. The 200-mb streamline charts have not been reproduced since the flow was straight easterlies and lack of adequate data prevents proper isotach analysis of the flow patterns.

The passage of the jet over Calcutta may, however, be seen clearly in Fig. 1(b). There was a rapid rise of wind speed at the 150-mb level from 100°, 32 knots at 0300 GMT of 2 September to 080°, 72 knots at 0300 GMT of 3 September. The surface trough intensified into a depression in the northeast Bay centred at 0300 GMT of 2 September near Lat. 21°N, Long. 90°E. It rapidly intensified into a cyclonic storm by the evening of the same day and moved almost westwards towards the coast near Balasore by the morning of 3rd. It is clear from the time-section in Fig. 1(b), that the development was in association with the jet maximum, since there is no evidence for the passage of a wave in the upper troposphere after the extension of the sea level trough into the northeast Bay. Lack of high level wind data does not permit the exact determination of the time when the jet maximum passed over Calcutta. By 1500 GMT of 4 September, the jet maximum should, however, have passed over Calcutta, since between 0300 GMT of 3rd to 1500 GMT of 4th, the 200-mb winds dropped from 100°, 58 knots to 090°, 22 knots. This would indicate that the cyclone was under the influence of the rear of the jet maximum during this period, when it weakened. The cyclone crossed the coast on the evening of 3rd, weakened into a depression and moved away westnorthwestwards.

5. Mechanism for the development of monsoon depressions

The hypothesis evolved at the end of Section 3 indicates that the deciding factor for the development of the sea level trough into a monsoon depression is the approach of the upper trough or wind maximum. It has been shown in the sub-section 4·1 that development occurs when the forward portion of the upper wave trough is superimposed over the sea level trough. When air streams ahead through the trough, it loses cyclonic vorticity and hence is subject to divergence according to the vorticity theorem. A zone of positive vorticity advection (vorticity diminishing downstream) lies ahead of the high level easterly trough and negative vorticity advection in its rear. This vorticity advection is given by

$$A_Q = -V\frac{\partial Q}{\partial s}$$

where V, the wind velocity and s, the distance are measured along the streamlines. Expressing absolute vorticity Q in terms of wind speed and direction,

$$Q = VK_s - \frac{\partial V}{\partial n} + f \qquad (2)$$

Petterssen (1955) has shown that neglecting the variation of shear along the streamlines,

$$A_Q = -V^2\left(\frac{\partial K_s}{\partial s} + K_s K_n\right) \qquad (3)$$

where K_s and K_n are the streamline and orthogonal curvatures respectively and n is taken normal to the streamlines.

It is seen that the vorticity advection, being proportional to the square of the wind speed, should be large in the upper troposphere, where strong winds are involved.

The second case in which development occurs even more rapidly than in the previous type, is with the approach of a wind maximum over Calcutta. In the case of a jet maximum of this type, it will not be permissible to neglect the variation of shear along the streamlines. The vorticity advection is expressed in full by the equation,

$$A_Q = -V^2\left(\frac{\partial K_s}{\partial s} + K_s K_n\right) + V\frac{\partial}{\partial s}\left(\frac{\partial V}{\partial n}\right) \qquad (4)$$

In the case of straight currents observed with easterly jets, the curvature is nearly zero, and the first term may not be important. The shear term on the cyclonic side of the jet may be as high as 3 or 4 times the Coriolis parameter or even more, while on the anticyclonic side, it does not generally exceed f (Koteswaram 1956). Riehl, Berry and Maynard (1955) found a proportionality between the absolute vorticity on the cyclonic side and the speed of some westerly jet streams over the U.S.A., the absolute vorticity ranging from 2 to 8×10^{-4} sec^{-1}. In the case of short jetlets, the variation of shear downstream towards the exit is also considerable and taking into account the high value of V, the value of vorticity advection is likely to be quite large in the left forward sector of a jet maximum. In the right rear sector, where there is also positive vorticity advection, it is much less, due to the smaller variation of shear downstream. Bundgaard (1956) has computed the vorticity advection in the case of typical jet streams and finds maximum positive advection in the left forward sector of straight jets.

It is now clear that in both the situations mentioned in the hypothesis in Section 3, there is large positive vorticity advection and associated divergence in the upper troposphere and it should be compensated by a corresponding convergence and generation of vorticity in the lower troposphere. The vorticity advection term A_Q in Petterssen's equation of development (Eq. 1), therefore, assumes major importance in the process of initial development of the surface vortex. Negative isallobaric centres referred to as 'low pressure waves' are known to move from the east and cause unsettled conditions or extend the sea level trough over the north Bay. These appear to be associated with the vorticity advection in the upper troposphere. With the sea level trough extending into the north Bay, a field of cyclonic vorticity at the sea level is created and later built up into a depression or storm with subsequent passage of an upper tropospheric perturbation and associated positive vorticity advection. The effect of upper divergence associated with westerly jet stream waves in the pre-monsoon season has been studied by Ramaswamy (1956) who atrributes the formation of convective thunderstorms and duststorms over northern India to the positive vorticity advection associated with the eastward passage of such waves in the upper troposphere.

The role of the other terms involving thermal factors in the equation (1) may now be examined.

(*i*) *Non-adiabatic heating* — Petterssen (1956) has pointed out that if the atmosphere were at rest relative to the earth, equation (1) would reduce to

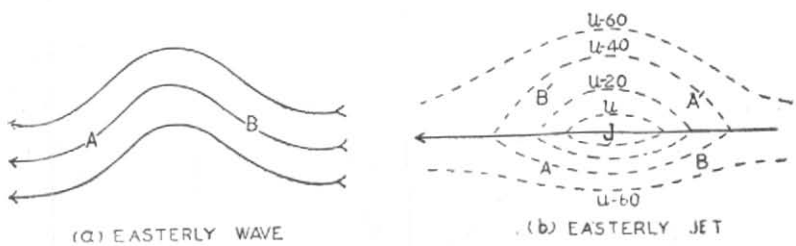
$$\frac{\partial q_o}{\partial t} = -\frac{R}{f}\nabla^2 H \qquad (5)$$

Where q_o is the relative vorticity at sea level, and

$$H = \log\left(\frac{p_o}{p}\right)\frac{1}{C_p}\cdot\frac{\partial \overline{W}}{\partial t},$$

where $\partial W/\partial t$ is the heat (other than latent) supplied to (or removed from) a unit mass per unit time, the bar denotes mean values through the layer from p_o to p (1000 mb to the level of non-divergence) and other symbols have their conventional meanings. Equation (5) indicates the effect of non-adiabatic heating or cooling on the development of vorticity. With the monsoon trough extending into the north Bay, fresh monsoon air, which is generally colder than the stagnant monsoon, advances from the southern parts of the Bay into the north Bay. It gains in temperature from the underlying surface as it moves north and as such H is positive, $\nabla^2 H$ is negative and the contribution of this term is positive. It is well known that Bay depressions close to coast are found to become more diffuse and the centre is even seen to cross coast during afternoon and shift back into the Bay during the night, except when the heavy rain occurring all round the depression reduces the thermal contrast between land and sea. This is apparently due to the non-adiabatic effect produced by equation (5).

(*ii*) *Adiabatic influences*—This is represented by the stability or buoyancy term $\overline{-R\nabla^2 S/f}$, where $S = \log(p_o/p)$. $\omega(\Gamma_a - \Gamma)$, where ω is the vertical velocity and Γ_a and Γ are the adiabatic (dry or wet) and actual lapse rates expressed in terms of pressure. ω is negative for upward motion and positive for downward motion. When the monsoon trough extends into the Bay, there is upward motion as evidenced by the occurrence of weather and the actual lapse rate is greater than the moist adiabatic lapse rate. This term also, therefore, contributes positively to development.



Figs. 3(a) and 3(b). Models of upper tropospheric perturbations leading to cyclogenesis in the north Bay during the monsoon season

A—Area of positive vorticity advection favourable for low level cyclogenesis

B—Area of negative vorticity advection unfavourable for low level cyclogenesis

In the case of easterly jet, A>A′ and B>B′

Streamlines are represented by continuous lines while broken lines represent isotachs; u indicates maximum wind speed in knots

(*iii*) *Thermal advection*—The third thermal factor involved is the thermal advection represented by the term

$$-\frac{g}{f}\ \nabla^2 A_T$$

where A_T denotes the effect of advection of thickness. During the stage of formation of the monsoon depressions, no large thickness advection is involved. After the formation of the depression, however, a thickness high ahead of the centre becomes well-marked and there is a positive contribution to vorticity in advance of the depression as in the case of extra-tropical cyclones. This term, therefore, contributes to the movement of the depression, since the centre should move from falling to rising vorticity tendencies. Such a tendency has been noticed by George (1953) in a study of a few tropical cyclones in the Bay of Bengal in the pre- and post-monsoon seasons. A detailed study of the vertical structure of the monsoon depression is in progress and the contributions of the various terms in the intensification and movement of the disturbances will be discussed in a subsequent paper.

It is seen from the above analysis, that apart from the upper vorticity advection, all the other terms in the development equation contribute positively to cyclonic development in the north Bay of Bengal during the southwest monsoon season. Since every upper wave does not produce lower cyclonic development, the thermal terms must be acting as breaks on occasions of non-development. Under such circumstances, *i.e.*, when the monsoon trough does not extend into the north Bay, westerly air from the north Peninsula moves into the north Bay from a warmer to a cooler region. As such the non-adiabatic term has a negative effect on the development of vorticity. No appreciable weather occurs over the north Bay and hence there is little or no upward motion : the buoyancy term does not make any positive contribution. Further, if there is any local subsidence the contribution would even be negative. The thermal advection term also does not make any appreciable contribution, since there is very little thickness gradient.

The part played by the lower trough in removing the obstacles in vorticity development at sea level initiated by the upper troughs of jets can now be understood. The lower trough provides the mechanism for the positive vorticity contribution by thermal influences.

6. Conclusion

The present investigation leads to the following hypothesis for the formation of monsoon depressions in the Bay of Bengal, *viz.*, cyclonic development at sea level occurs when and where an area of positive vorticity advection in the upper troposphere becomes superimposed upon a pre-existing trough at sea level. The similarity of this mechanism with that found by Petterssen (1955) for extra-tropical cyclogenesis is striking.

Figs. 3(a) and 3(b) illustrate two typical types of upper vorticity advection—the easterly wave trough and the easterly jet maximum. Maximum positive vorticity advection would occur in the regions marked A and negative vorticity advection in the zones marked B. Maximum development would, therefore, occur when the sea level trough is located under A and the depression would weaken when it comes under B. Only the sinusoidal wave and the straight jet have been considered here. Vorticity advection would differ in location and intensity in the case of confluent or diffluent troughs and jet streams imbedded in wave troughs. Examples of such cases in westerly currents were given by Riehl (1954), Petterssen (1956) and Ramaswamy (1956).

The application of the above hypothesis to cases of cyclonic development over land, *i.e.*, when the monsoon trough is located over northern India, would point to the formation of land depressions. A few such cases have also been noticed in the present investigations. Land depressions also occasionally form under the influence of positive vorticity advection ahead of upper westerly wave troughs. Such cases occur generally before the incursion of the southwest monsoon over the area concerned or after its withdrawal, since during the period when the monsoon is active, the general high level flow is easterly and there is very little possibility of westerly waves in the upper troposphere.

REFERENCES

Bjerknes, J.	1937	*Met. Z.*, **54**.
	1951	*Compendium of Meteorology*, p. 577.
Bundgaard, R. C.	1956	*Bull. Amer. met. Soc.*, **37**, 9, p. 465.
Desai, B. N.	1950	*Mem. India met. Dep.*, **28**, pt. 5.
Desai, B. N. and Koteswaram, P.	1951	*Indian J. Met. Geophys.*, **2**, 3, p. 250.
Dines, W. H.	1912	*Geophys. Mem.*, 2, Met. Office, London.
Eliot, John	1900	*Handbook of Cyclonic Storms*, India met. Dep., **1**, p. 137.
George, C. A.	1953	*Indian J. Met. Geophys.*, **4**, 4, p. 279.
Koteswaram, P.	1956	Dep. Met. Univ. Chicago, *Res. Rep.*, May 1956.
Malurkar, S. L.	1950	*Mem. India met. Dep.*, **28**, pt. 4, p. 160.
Mull, S. and Rao, Y. P.	1949	*Indian J. Phys.*, **23**, 40, p. 371.
Palmer, C. E.	1951	*Compendium of Meteorology*, p. 859.
Petterssen, S.	1955	*J. Met.*, **12**, p. 36.
	1956	*Weather Analysis and Forecasting*, 2nd Edition, **1**, Motion and Motion System, McGraw Hill, New York.
Petterssen, S., Dunn, G. E. and Means, L. L.	1955	*J. Met.*, **12**, 1, p. 58.
Pisharoty, P. R. and Kulkarni, S. B.	1956	*Indian J. Met. Geophys.*, **7**, 2, p. 103.
Pramanik, S. K. and Rao, Y. P.	1948	*Sci. & Cult.*, **14**, 1, p. 34.
Ramanathan, K. R. and Ramakrishnan, K. P.	1932	*Mem. India met. Dep.*, **26**, pt. 2, p. 13.
Ramaswamy, C.	1956	*Tellus*, **8**, 1, p. 26.
Riehl, H.	1945	Dep. Met., Univ. Chicago, *Misc. Rep.*, No. 17.
	1948	*Ibid.*, No. 24.
	1954(a)	*Met. monograph*, Amer. met. Soc., **2**.
	1954(b)	*Tropical Meteorology*, McGraw Hill, New York.
Riehl, H., Berry, F. A. and Maynard, H.	1955	*J. Met.*, **12**, 1, p. 26.
Roy, A. K.	1946	*India met. Dep. Tech. Note*, 16.
Roy, S. C. and Roy, A. K.	1930	*Beitr. Phys.*, **16**, p. 234.
Sur, N. K.	1932	*Mem. India met. Dep.*, **26**, pt. 2.
Sutcliffe, R. C.	1947	*Quart. J.R. met. Soc.*, **73**, p. 310.
Sutcliffe, R. C. and Forsdyke, A. G.	1951	*Ibid.*, **76**, p. 189.