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Upper air contour patterns and associated heavy rainfall during the southwest monsoon

P. R. PISHAROTY and S. B. KULKARNI

Meteorological Office, Poona

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ABSTRACT. The necessity of dealing with the lower troposphere over India as consisting of two separate strata having opposite values of horizontal wind divergence is briefly explained. It may not be sufficient to look for areas of convergence at the lower levels; areas of marked divergence at the upper levels of 500 mb and aloft may be equally, if not more, important for the understanding and forecasting of very heavy rains. The principles of deducing areas of marked horizontal divergence on qualitative reasoning based on considerations of non-geostrophic motion as well as of air flow down the gradients of vorticity are then outlined. A few cases of non-orographic phenomenal rainfall of 10″-20″ in 24 hours which fell over certain parts of India during the southwest monsoon of 1954 are shown to have fallen in areas of marked upper air divergence, deducible from such qualitative reasoning applied to flow patterns at 500 mb and aloft.

1. Introduction

It is well known that the *southwest* sector of depressions and cyclones of the southwest monsoon season is usually the region of heavy rainfall of 3″ and above in twentyfour hours. There are various explanations frontal and otherwise for this phenomenon (Ramanathan and Ramakrishnan 1933, Desai 1951, Mull and Rao 1949). However, every Indian forecaster knows that there are numerous occasions during the southwest monsoon season, when other sectors of such cyclonic systems experience heavy rainfall. There are also occasions of heavy rain which are not associated with westward moving cyclonic systems detectable on the sea level charts. We have no accepted criteria to foresee or even account for the heavy rainfall in sectors other than the southwest sector of monsoon depressions.

The paper describes a study in which a few occasions of heavy rainfall over the Bombay State are examined on the basis of day-to-day 500-mb contour patterns and the 700-500 mb thickness lines. The thickness patterns were drawn with the help of pibal winds at 0900 GMT and the thickness values at the dozen radiosonde observations taken at 1500 GMT, and are therefore liable to a certain amount of criticism. The 500-mb contour patterns were built up from the surface charts through the 1000-mb chart, and the thickness patterns between 700 and 1000-mb levels and those between the 500 and 700-mb levels. It was encouraging to find that quite a few of the occasions of heavy rainfall, even when they did not occur over the southwest sector of the westward moving cyclonic systems of the sea level chart, were associated with a contemporaneous flow pattern at the 500-mb level, which suggested marked horizontal divergence on grounds of qualitative dynamic reasoning.

2. Convergence and divergence in the atmosphere

Computations of convergence on the basis of actual pibal winds in the first 10,000 ft of the atmosphere over India have been carried out by Das (1951) and Asnani (1951) for specific occasions. They have obtained convergence values varying between 3×10^{-5} per sec and 9×10^{-5} per sec for levels between 5000 and 10,000 ft a.s.l. The contribution to the rate of pressure change at the ground level due to the convergence in the layers of the atmosphere between the surface and a height level z, can be computed on the assumption that there is no horizontal divergence in the superincumbent column of air. According to the partial tendency equation, the surface tendency of pressure is given by

$$\frac{\partial p_0}{\partial t} = -g \int_0^z \rho \operatorname{div}_z \mathbf{v} \, dz \qquad (1)$$

where g is the acceleration due to gravity, ρ is the local density of air and $\operatorname{div}_z \mathbf{v}$ is the horizontal wind divergence. Assuming a mean value of the horizontal wind divergence of 3×10^{-5} per sec between the 1000 and 700-mb levels (approximately the lowest 10,000 ft of the atmosphere), this surface pressure tendency would amount to 9 dynes cm^{-2} per sec or about 800 mb in 24 hours.

This is a colossal amount. The 24-hour pressure changes in tropical latitudes rarely exceed 5 mb even on days of heavy rain. Upward motion through a level surface at 10,000 ft or so will not decrease the magnitude of this surface pressure change, as long as there is no appreciable horizontal wind divergence in the superincumbent layers of the atmosphere. As originally emphasized by Bjerknes and Holmboe (1944) we are forced to conclude that the atmosphere must be analysed as composed of two strata having opposite values of horizontal wind divergence of roughly equal numerical values. These strata are necessarily separated by a level of non-divergence whose height may vary from one synoptic situation to another.

We are not yet in a position to state definitely the average height of the non-divergent layer over the Indian area. But since horizontal wind convergence has been noticed at the 10,000 ft level only on relatively rare occasions, we may tentatively take the mean height of this level to be 10,000 ft or above. General studies over the middle latitudes have revealed this level of non-divergence, and therefore of approximate barotropic motion, to be near the 600-mb level.

Hence a proper analysis of the Indian atmosphere should consist of an examination of two strata, one preferably at the 850-mb level (or the 5000-ft level), and the other at the 500-mb level or even better at the 300-mb level. The well-known forecasting rule of heavy rains in the southwest sector, is apparently based on the studies of the commonly occurring flow patterns leading to convergence in lower strata. As the pibal ascents are absent on cloudy days and therefore cannot provide an idea of the contemporaneous flow patterns in the upper levels, our attention was of necessity confined to the lower levels, and the current forecasting rules are based on the low level flow patterns.

3. Flow patterns and associated divergence (without taking the vorticity field into consideration)

It is possible to obtain qualitative ideas of areas of divergence from a knowledge of the flow patterns when streamlines and isotachs are drawn with sufficient accuracy. These have been described by Bjerknes and Sandstorm (1911) and more recently by Palmer (1952) and Riehl (1954). If the isotachs show an increase of wind velocity downwind in a region of parallel streamlines, it is an area of divergence and *vice versa*. However, the divergence does not become kinematically obvious when there is a decrease of wind speed downwind and a simultaneous diffluence of the streamlines.

In the case of geostrophic motion on a rotating earth, the upper air contours become the streamlines and the magnitude of the wind is inversely proportional to the separation between the contours. In these circumstances, the mass transported through any contour channel is independent of the diffluence or confluence of contours, provided there is no latitude change*. Hence the well known principle, that under strict latitudinal geostrophic motion there is no divergence associated with a diffluent contour pattern. However, we have to examine the mechanism by which a strong wind at a constriction of the contours can become weaker with the diffluence of the contours. On a rotating earth,

*This assumption is only to avoid, in this part of the discussion, the complication introduced by the changes in the Coriolis force with changes in latitude in the frictionless free atmosphere, all the forces act normal to the wind, so that a geostrophic or even a gradient wind cannot decelerate or accelerate. Hence strict geostrophic motion is not possible whenever there is a diffluence or confluence of patterns, even when the flow is latitudinal. Elementary reasoning, given below, shows that a weakening of the wind in an area of diffluent contours can take place only by the development of a non-geostrophic component across the contours towards the region of higher values.

Let the geostrophic wind at A be V (Fig. 1). At this point, the pressure gradient force AP balances the Coriolis force AQ, corresponding to the velocity V and the latitude of the point A. When the particle of air reaches a point B a distance dx downstream from A, the velocity remains the same as there are no forces acting against the flow, except friction, which we neglect in the free air. Hence the Coriolis force BQ′ remains equal to AQ. However, due to the greater separation between the contours, the pressure gradient force BP′ is less than what it was at A. Consequently balanced motion is not possible and particles of air will be deflected across the contours towards the higher contour. It is this mechanism which would decrease the wind speed down to the value appropriate to the geostrophic value in due course. The cross-isobar component will be removing air from the channel at its southern boundary but feeding into channel at the northern boundary. If the degree of diffluence varies across the isobars, there may be an excess of the feed or removal of air on account of the cross-isobar component. It is easy to see that decreasing diffluence towards decreasing pressure will cause divergence and increasing diffluence will cause convergence.

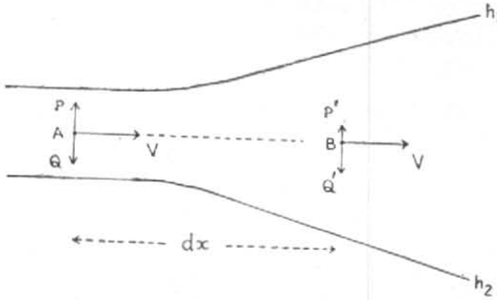

Fig. 1. Development of an ageostrophic component in a region of diffluent contour

4. Deduction of areas of divergence from vorticity considerations

Neglecting terms arising from vertical motion and from horizontal temperature gradients, Rossby has deduced the vorticity equation for large scale atmospheric flow, in the form—

$$\frac{d\zeta_a}{dt} + \zeta_a \operatorname{div}_2 \mathbf{v} = 0 \qquad (2)$$

where ζ_a is the vertical component of the absolute vorticity of a parcel of air and $\operatorname{div}_2\mathbf{v}$ is the horizontal wind divergence.

If the relative vorticity with respect to the earth be denoted by ζ, the absolute vorticity to be used in the equation is given by $\zeta + l$ where l is the Coriolis parameter $2\Omega \sin \phi$. Rewriting equation (2) we get—

$$\operatorname{div}_2 \mathbf{v} = -\frac{1}{\zeta_a}\left(\frac{\partial\zeta_a}{\partial t} + V\frac{\partial\zeta_a}{\partial s}\right) \qquad (3)$$

where $\frac{\partial\zeta_a}{\partial t}$ is the local rate of change of the absolute vorticity about a vertical axis and $V\frac{\partial\zeta_a}{\partial s}$ is the advective rate of change of the same quantity.

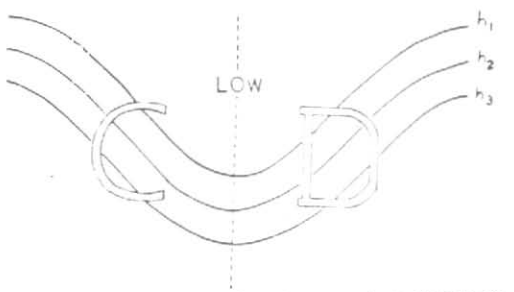

Fig. 2. Cyclonic vorticity due to curvature decreases downstream in the region D (divergence) and increases downstream in the region C (convergence)

In many situations, $\frac{\partial \zeta_a}{\partial t}$ may be small in comparison with $V\frac{\partial \zeta_a}{\partial s}$ so that as a first approximation we may write

$$\mathrm{div}_2\ \mathbf{v} = -\frac{1}{\zeta_a}\ V.\frac{\partial \zeta_a}{\partial s} \tag{4}$$

Further, if the changes in the total vorticity contributed by the changes in latitude be small, the expression further simplifies to—

$$\mathrm{div}_2\ \mathbf{v} = -\frac{1}{\zeta_a}\ V.\ \frac{\partial \zeta}{\partial s} \tag{5}$$

where ζ is the relative vorticity with respect to the earth.

Equation (5) really means that practically the whole magnitude of the horizontal divergence is required to bring about the necessary change in the vorticity advected downstream.*

The relative vorticity ζ is composed of the vorticity due to the shear and the vorticity due to curvature and can be expressed by

$$\zeta = \left(\frac{V}{r} - \frac{\partial V}{\partial n}\right)$$

where r is the cyclonic radius of curvature and n is the normal drawn towards the left of the direction of the wind in the northern hemisphere.

While the geostrophic winds are not good enough for computing divergence values, they have often been found to be good enough for estimating vorticities. The calculation of vorticity gradients from these values of vorticity is a tricky affair, and can be done only when the gradients are conspicuously large. Hence, in such cases, it is possible to estimate the sign of $\frac{\partial}{\partial s}\left(\frac{V}{r} - \frac{\partial V}{\partial n}\right)$ from the contour patterns, and thereby that of the horizontal divergence.** In other words regions where cyclonic vorticity decreases downstream may be regarded as regions of horizontal divergence (Riehl 1952, 1954).

An example of change of cyclonic vorticity downstream is given in Fig. 2.

In Fig. 2, h_1, h_2, h_3 represent three contours on the 500-mb chart, the spacing between them remaining practically constant. Neglecting changes in latitude and changes in centrifugal accelerations, the winds remain the same all along the contour channels. However, the changes in the curvature of the contours from a cyclonic bend at the trough line to an anticyclonic curvature downstream through the region of straight contours, makes D a region in which cyclonic vorticity is decreasing. If winds are to blow through the contour channels of this pattern, the region D should be an area of horizontal divergence even if there is no diffluence of the contours. Similarly, the region C should be a region of horizontal convergence.

*As shown by Bjerknes (1951) this can be so only as long as the vertical component of the absolute vorticity is large. The above simplication will cease to be valid, when ζ_a becomes small, *i.e.*,in areas downstream from sharply curved anticyclonic bends.

**It is very tempting to devise numerical methods by which these vorticity gradients may be computed from contour fields. When the gradients are large enough, they can be deduced qualitatively without a computation. But when they are small as to need a computation, the geostrophic winds, particularly over the Indian latitudes, are not good enough to give the *vorticity gradients* with sufficient accuracy. This is a point often forgotten by some workers.

In actual patterns observed on a synoptic 500-mb chart, there will be areas where the cyclonic vorticity will be due to curvature and shear effects. There will be regions where the curvature and shear effects co-operate in giving a gradient downstream. In such cases qualitative deductions about areas of divergence would be justified. But when the shear and the curvature effects produce opposite gradients in an area, one cannot make reliable qualitative deductions about divergence. In such cases of opposite gradients, quantitative deductions about the total effects are equally unreliable unless they are deduced from a close network of *actual* wind observations.

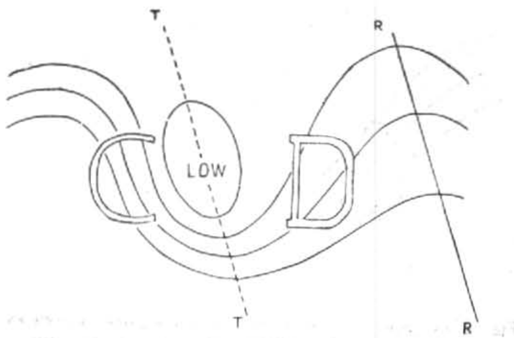

Fig. 3

Region D is an area of diffluence of contours where strong winds developed in the region C weaken appreciably. It is also an area where marked cyclonic vorticity at the trough line TT changes over to anticyclonic vorticity at the ridge line RR. As the diffluence commences at some distance to the left of the trough line TT, the area of divergence D really extends slightly to the left of the trough line TT

5. Flow patterns having diffluence and decrease of vorticity downstream

There can be situations in which an area on the 500-mb chart shows diffluence of the contours as well as decrease of cyclonic vorticity downstream. Such areas should, therefore, be areas of marked horizontal divergence at this level. Fig. 3 is a schematic diagram illustrative of the point under discussion.

The diffluent trough discussed by Bjerknes (1954) is practically the same as what is illustrated in Fig. 3.

In arriving at firm conclusions regarding divergence in areas similar to that at D in Fig. 3, it is necessary to examine the patterns at a higher level, *e.g.*, at 300-mb level also. If the area to the left of the trough line at 500-mb is colder, the trough line will shift to the left with height, or will slope westwards, so that the area D will be to the east of the trough line, at 300 mb. In such a situation, the flow pattern at the 300-mb level also calls for divergence above the area which has divergence at the 500-mb level thereby increasing the total divergence in the stratum 500-300 mb standing over the area D. This would call for increased compensating convergence in the lower strata. On the other hand, if the trough line at 300-mb level were to the east of the trough line at 500 mb, the total divergence in the column between 500 and 300 mb would be less and so also the compensating convergence in the lower stratum.

6. The development equation of Sutcliffe and deduction of divergence from thermal patterns

Sutcliffe (1947) has derived an expression for the excess of isobaric divergence at a higher level, over that of the isobaric divergence at a lower level, in terms of absolute vorticity fields and thermal winds. Since we are usually applying data on constant pressure surfaces for analysis, Sutcliffe's expressions are theoretically more justified. In Indian latitudes, where the inclination between the constant pressure surfaces and the horizontal are much smaller than what prevails in higher latitudes, the refinement in Sutcliffe's expressions may be, to some extent, superfluous for the Indian area. According to Sutcliffe the excess of the isobaric wind divergence at the higher level (p), over that at a lower level (o) is given by

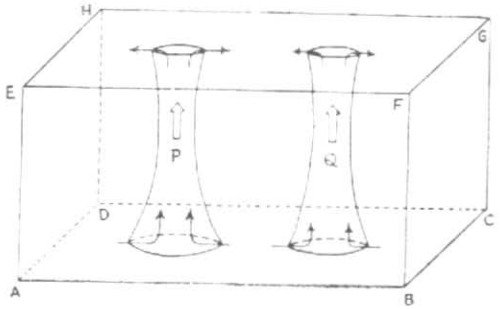

Fig. 4

ABCD is a level where there is marked convergence —say the 850-mb level. EFGH is a higher level where there is marked divergence—say the 500-mb level. The vertical transport of air from the lower level to the higher level may be effected either by a gentle vertical motion throughout a cross-section equal in area to either ABCD or EFGH, or through narrow columns represented schematically by P and Q

$$(\mathrm{div}_p\, \mathbf{v} - \mathrm{div}_o\, \mathbf{v}_o) = \frac{-\mathbf{v}_T}{2\,\Omega\,\sin\phi} \cdot \frac{\partial}{\partial s}\,(2\,\Omega\,\sin\phi + 2\zeta_o + \zeta_T) \qquad (6)$$

where,

$\mathrm{div}_p\, \mathbf{v}$ = the isobaric wind divergence at the higher level p,

$\mathrm{div}_o\, \mathbf{v}_o$ = the isobaric wind divergence at the lower level o.

$\mathbf{v}_T$ = the thermal wind, or wind shear between the higher and the lower levels,

ζ_T = the vertical component of the vorticity of the thermal wind,

ζ_o = the vertical component of the relative vorticity at the lower level and

$\frac{\partial}{\partial s}$ = the differentiation along the direction of the thermal wind.

In a few publications that have appeared in the recent past, computations of this relative divergence have been carried out for the atmospheric layer between 1000 and 500-mb levels, and the development of cloudiness and rainfall associated with this relative level divergence. From what we have seen in Section 2, it is clear that we should analyse the processes in the atmosphere in two separate strata of the lower troposphere. If there is predominent divergence in the lower of these two strata, or if there is divergence in both of these two strata, physical reasoning precludes developments of upward motion and consequent clouds and precipitation. Sutcliffe's expression if used separately for the two strata 1000–700 mb and 700–500 mb will bring out the magnitudes of the lower level convergence and the upper level divergence and thereby the possibility of development of weather or otherwise. An examination of the changes in the vorticity revealed by the thermal wind field between 500 and 700-mb level will be a guide for determining the upper level divergence which when conspicuous is also likely to be associated with an equally large convergence in the lower stratum.

7. Upward motion and divergence areas

If an area of marked divergence at the upper levels is superposed over an area of marked convergence at the lower levels, it is reasonable to expect an upward transport of air from the lower to the higher levels. Consequently, clouds and precipitation can develop over such an area, if the moisture conditions in the lower stratum are favourable. It is noticed from synoptic charts that even over areas of widespread rain exceptionally heavy rainfalls are highly localised over much smaller areas. There is probably much greater convergence over such small areas of exceptionally heavy rain compared to the surroundings. Such localised convergence in relatively narrow columns is schematically represented in Fig. 4. Convergence computed from an open network of pibal or radiosonde stations cannot be expected to locate these small areas of relatively large convergence.

8. Application to forecasting during the southwest monsoon

Actual computations of the various terms in equations (4), (5) and (6) have been carried out for specific synoptic situations by numerous authors. With the present staff position, such computations are not likely to be adopted on a routine basis, at the various Forecasting Offices in India. Here we would also like to add that such numerical computations are not capable of giving reliable results on days when the divergence over any area under examination is not conspicuously large enough to be obvious on qualitative reasoning. While non-geostrophic components may not vitiate the general vorticity values, these non-geostrophic components would seriously effect the gradients of vorticity on days of weak vorticity gradients so that computations based on geostrophic winds would yield spurious results on such occasions.

Hence the aim in this paper has been to locate the areas of conspicuous divergence by qualitative reasoning based on the aspects of diffluence and vorticity gradients. It is true that there are occasions when the shear terms and the curvature terms produce opposite gradients of vorticity in a particular area so that evaluation of the divergence by qualitative reasoning is bound to be faulty. But when the shear terms and the curvature terms distinctly act in the same sense as far as vorticity gradients are concerned, the regions are bound to be regions of large horizontal divergence. When this feature occurs during the monsoon in regions where there is an abundant supply of moisture in the lower levels, such areas are likely to be areas of very heavy rains. In this paper, there has been no attempt to forecast such areas of divergence. The aim has been to examine whether such areas of obvious upper air divergence, as revealed by the contour patterns and the thickness patterns, are areas of contemporaneous heavy rainfall. The examination has been encouraging. It is not claimed that all regions of heavy rainfall are invariably revealed by these upper air patterns, in the manner described. The point we wish to make out is only that these regions are worth looking for, in explaining regions of heavy rainfall during the monsoon period.

9. A few examples of phenomenal rainfall during July—September 1954

1 July 1954

By the morning of 1 July, the sea level trough which was present in the northwest Arabian Sea on 30 June had weakened and the consequent release of moist winds caused 3″ of rain in 24 hrs at Surat and widespread light rain over Saurashtra-Kutch.

The constant pressure charts corresponding to 1500 GMT of 1 July reveal many interesting features over the Gulf of Cambay and the adjoining areas (Fig. 5). The 700–500 mb thickness lines show a trough extending from Rajasthan to Saurashtra, with its axis passing through Bikaner and Veraval. The 500-mb contours also show a similar trough with a region of marked diffluence over north Konkan, south Gujarat and adjoining districts of Madhya Bharat. At the 300-mb level, the axis of the trough was roughly along Long. 70°E, *i.e.*, somewhat to the west of the trough at the 500-mb level. From the reasoning outlined in Sections 3, 4, 5 and 6, it is clear that the situation demands marked upper air divergence above the 700-mb level, over north Konkan, south Gujarat and adjoining districts of Madhya Bharat. Large upward motions commencing in the moist air below and extending to great heights can occur in this area of marked diffluence. As explained in Section 7, this upward motion may be highly localised even within this area. Very heavy falls of rain were reported on the 2nd morning—18″ from Bulsar, 15″ from Panvel; and 17″ from Kathiawara in Alirajpur district of south Madhya Bharat. It is also reported that of the 17″ at Kathiawara 15″ fell within 5 hours.

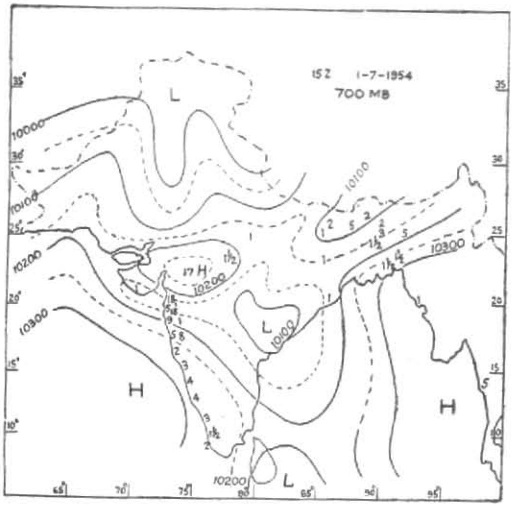

Fig. 5(a)

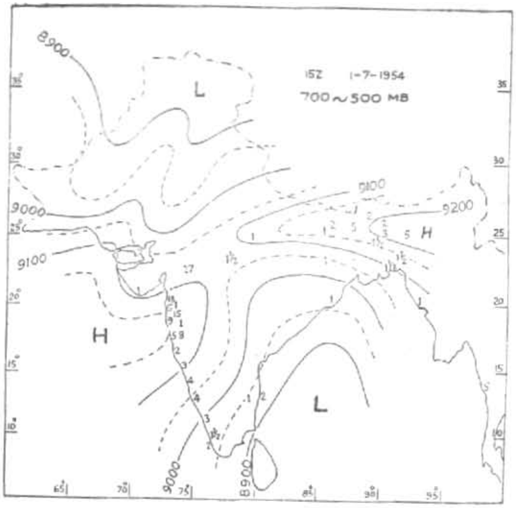

Fig. 5(b)

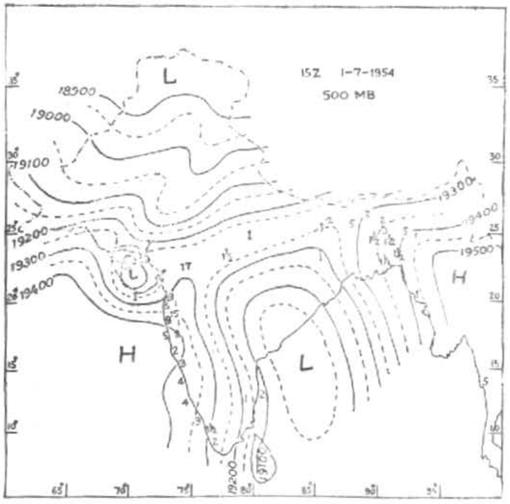

Fig. 5(c)

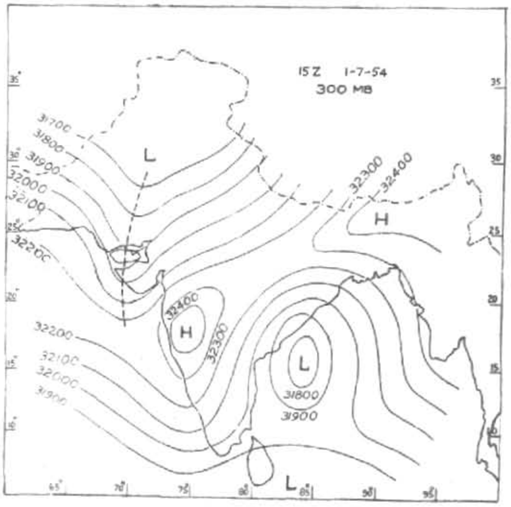

Fig. 5(d)

Fig. 5. Contours and thicknesses at 15Z of 1 July 1954 in units of geopotential feet

(The rainfall amounts of one inch and more, recorded for the 24 hours ending 03 Z of 2 July are entered at the appropriate places)

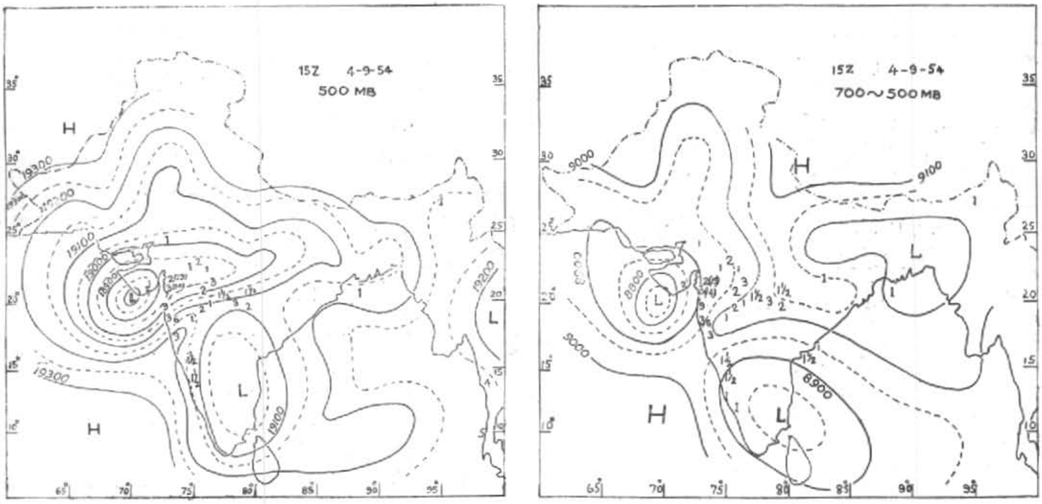

Fig. 6(a) Fig. 6(b)

Fig. 6. Contours and thicknesses at 15Z of 4 September 1954 in units of geopotential feet

(The rainfall amounts of one inch and more recorded during the 24 hours ending 03 Z of 5 September are entered at the appropriate places. The figures in brackets are the rainfall amounts recorded during the 24 hours ending 03 Z of 6 September)

4 September 1954

On the 4th morning a depression lay over east Madhya Pradesh centred near Gondia. The 500-mb chart of 1500 GMT of the 4th was very interesting in that Veraval, which showed a lower height than Bombay and Karachi even at 1500 GMT of 3rd, showed a further fall of 260 ft. The actual height at Veraval on the 4th was 18,800 ft, nearly 400 ft below that at Poona and 300 ft below that at Karachi (Fig. 6). Consequently there was a deep low centred close to Veraval with a markedly diffluent area over south Gujarat and adjoining north Konkan. (We would have completely missed this low, if we had no radiosonde data from Veraval or if we had ignored it). The reasoning outlined in *Section 4* calls for marked divergence over south Gujarat and adjoining north Konkan. The depression over Madhya Pradesh had strengthened the monsoon over the Konkan coast. By the afternoon of the 4th, Dahanu had 7·3″ of rain which became 9·4″ by the 5th morning. By the 5th evening Surat had 14″ rain and Broach 4″, all recorded since that morning. By the 6th morning the India Meteorological Department observatory at Broach recorded 12″ while the State Raingauge Station there recorded 19″ of rain. The deep 500-mb low over Veraval and the divergence area ahead of it had played their parts. On the 5th evening the height of the 500-mb surface at Veraval rose by 390 ft.

10. Acknowledgement

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