PART II PREPRINTS

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PART II

AMERICAN METEOROLOGICAL SOCIETY

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Part II: Papers received after November 1, 1973

FOREWORD

The International Tropical Meteorology Meeting was held January 31 - February 7, 1974 in the City Hall (Conference Hall), Nairobi, Kenya. Part I of the Preprint Volume contains manuscripts furnished by the authors in advance of the conference (prior to November 1, 1973); it was distributed to all registrants at the meeting.

Part II of the Preprint Volume contains manuscripts received after November 1, 1973. Also included in Part II are discussions held during the sessions.

Copies of the Preprint Volume, Parts I and II, may be purchased from the American Meteorological Society, 45 Beacon Street, Boston, Mass. 02108, USA.

PART II

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There are a few papers that were presented at the Meeting but are not included in Parts I and II of the Preprint Volume. These manuscripts were not available.

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COMPUTATION OF TEMPERATURE AT THE LIFTED CONDENSATION LEVEL

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1. INTRODUCTION

For its importance in the thermodynamics of the atmosphere, the temperature at the lifted condensation level has been studied by many authors. Haurwitz (1941) and Petterssen (1956) used approximate formulae for the dew point and dry bulb temperature changes of the lifted parcel, while Prosser and Foster (1966) and Stackpole (1967) used iteration processes. Barnes (1968) applied a statistical method. Saito (1968) used the hydrostatic equation and some other approximations. Inman (1969) approximated a power series in the dew point deficit. The formula by the latter author yields results accurate to within 0.05°C for dew point depression of 25°C or less. Saito[1] used iteration process in principle.

The purpose of this paper is to get without iteration technique more accurate formulae for the temperature at the lifted condensation level, even when the dew point depression is too large which is generally the case in the African Sahara, and also when the saturation water vapour pressure of the lifted parcel is considered over water and/or over ice during the lift.

2. APPROXIMATE FORMULAE

The lifted condensation level(LCL) is defined as the level at which a parcel of moist air would become saturated when lifted dry adiabatically with constant mixing ratio (Huscke, 1959). It is well known that the above mentioned definition does not agree with nature (Hess, 1959 and Petterssen, 1956) but it is usually used for its simplicity and will be applied in this paper.

From this definition, the temperature at the lifted condensation level can be given by solving the equations

$$\theta = T\,(P_o/P)^{R/C_p} \tag{1}$$

and

$$X = \epsilon\, e/(P - e) \tag{2}$$

together with Clausius-Clapeyron's equation

$$de/e = \epsilon L\, dT_d/RT_d^2 \tag{3}$$

where θ, T and T_d are the potential, dry bulb and dew point temperatures respectively in degrees kelvin at the initial pressure P, X is the humidity mixing ratio in grams per kilogram of dry air, e is the water vapour pressure, ϵ is the ratio between the densities of dry air and water vapour at the same temperature and under the same pressure and is equal to 0.622, R is the gas constant for moist air and will be used as equivalent to the gas constant for dry air R_d, C_p is the specific heat of the moist air under constant pressure and will be considered equal to the specific heat of dry air under constant pressure, L is the latent heat of vapourization (sublimation) and P_o is equal to 100cb.

From (1) and (2) and (3) it is easily found that

$$\ln(T_c/T) + \epsilon L/C_p\,(1/T_c - 1/T_d) = 0 \tag{4}$$

where T_c is the temperature at the LCL. In (4) the latent heat L is considered as a constant. When its change with temperature is taken into consideration, a similar formula can be obtained in the form:

$$\ln(T_c/T) + (\epsilon L(T_d)/C_p + \epsilon B T_d/C_p)\,(1/T_c - 1/T_d) + (\epsilon B/C_p)\ln(T_c/T_d) \tag{4'}$$

where

$$L_{vw}(T_d) = L_{vw}(T_o) - B(T_d - T_o)$$

and similarly (5)

$$L_{vi}(T_d) = L_{vi}(T_o) - B'(T_d - T_o) \quad \text{(near 0°C)},$$

$$B = C_w - C_{pv},\quad B' = C_i - C_{pv},$$

L_{vw} and L_{vi} are the latent heat for the change from water vapour to liquid water and from water vapour to ice respectively, T_o is equal to 273.16°C, C_{pv}, C_w and C_i are the specific heat of water vapour under constant pressure, the specific heat of water and ice respectively (see Holmboe, Forsythe and Gustin, 1948).

The equality:

$$\ln T_c = \ln T_d - \ln[1 + (T_d - T_c)/T_c] \tag{6}$$

is very suitable for solving (4) and (4'). In the meteorological ranges, the maximum difference between T_d and T_c is about 11°C corresponding to difference in T and T_d of about 100°C. Different forms of $\ln T_c$ are given in Appendix and for further details see Lasheen (1973).

Neglecting second and higher powers of $(T_d - T_c)/T_c$ in the expansion of $\ln[1 + (T_d - T_c)/T_c]$ in (6), then substituting in (4) and (4') respectively we get

$$T_c = T_d\,[\epsilon L/C_p T_d - 1]/[\epsilon L/C_p T_d - 1 + \ln(T/T_d)] \tag{7}$$

and

$$T_c = T_d\,[\epsilon L(T_d)/C_p T_d - 1]/[\epsilon \times L(T_d)/C_p T_d - 1 + \ln(T/T_d)]. \tag{7'}$$

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[1] Saito, N. Private communication, 1973.

The absolute error in T_c as given by (7) can be easily shown to satisfy the equation

$$T_{c2} - T_{c1} = -T_{c2}T_{c1}\mathcal{R}_2/[T_d(\epsilon L/C_pT_d) - 1] \quad (8)$$

where T_{c2} and T_{c1} are the values of T_c as given by (4) and (7) respectively. $\mathcal{R}_2$ is given by the equation

$$\ln T_{c2} = \ln T_d - [(T_d - T_{c2})/T_{c2} + \mathcal{R}_2]$$

and is equal to

$$-[(T_d - T_{c2})/T_{c2}]^2/2(1 + \zeta)^2.$$

ζ lies between 0 and $(T_d - T_{c2})/T_{c2}$.

It follows from (8) that

$$T_{c2} - T_{c1} < (T_d - T_{c2})^2/2T_d[(\epsilon L/C_pT_d) - 1](1+\zeta)^2$$
$$< (T_d - T_{c2})^2/2T_d[(\epsilon L/C_pT_d) - 1] \quad (8')$$

As mentioned before, the maximum difference between T_d and T_{c2} in the meteorological ranges is about 11°C which corresponds to difference between T and T_d of about 100°C. So (7) gives an underestimate of temperature at the LCL by less than 0.05°C for dew point depression of about 100°C.

Accepting that the saturation of water vapour at temperatures equal or less than -40°C is over ice, as assumed in the construction of Canada TEPHIGRAM (form No. 2355 Sept. 1952) then L_{vi} is used when $T_d \leqslant -40$°C. In some cases $T_d > -40$°C but T is relatively high and the decrease of T_d along the constant humidity mixing ratio curve is enough to make $T_d \leqslant -40$°C. So we have three different cases:

a) <u>Saturation water vapour pressure over water</u>

In this case the saturation water vapour pressure of the lifted parcel is considered over water only i.e.

$$T_d > -40°C \text{ and } T_c > -40°C$$

Using (1), (2) and (3) the above mentioned conditions can be rewritten in the form:

$$T_d > -40°C$$
$$A = \ln(T/233.16) + (\epsilon L_{vw}/C_p)(1/T_d - 1/233.16) < 0$$

We can apply (7) using the latent heat of vapourization in this case.

b) <u>Saturation water vapour pressure over ice</u>

In this case the saturation vapour pressure of the lifted parcel is considered over ice only i.e.

$$T_d \leqslant -40°C$$

Here we can use (7) with latent heat of sublimation.

c) <u>Saturation water vapour pressure over water and over ice</u>

In this case saturation water vapour pressure is considered over water whenever the dew point temperature of the lifted parcel is greater than -40°C, and over ice whenever its dew point temperature is equal or less than -40°C. This case occurs when

$$T_d > -40°C \text{ and } T_c \leqslant -40°C$$

Using (1) and (2) and (3) the above conditions can be written in the form:

$$T_d > -40°C \text{ and } A > 0$$

and the dry bulb temperature, the lifted parcel would attain when its dew point temperature becomes equal to -40°C can be given by

$$T^* = T\exp[(\epsilon L_{vw}/C_p)(1/T_d - 1/233.16).$$

After this we can apply (7) using the latent heat of sublimation and substituting T^* and 233.16 instead of T and T_d. Thus

$$T_c = 233.16[(\epsilon L_{vi}/233.16C_p) - 1]/[(\epsilon L_{vi}/233.16C_p) - 1 + \ln(T^*/233.16)] \quad (7^*)$$

(7*) can be written in the form:

$$T_c = 233.16[(E/233.16) - 1]/(D-1) \quad (9)$$

where

$$D = (E/233.16) + A,$$
$$E = (\epsilon L_{vi}/C_p).$$

It is worth mentioning that the application of (7) and (9) in meteorological ranges and classified cases as discussed above will reduce the maximum error previously discussed.

3. MORE ACCURATE FORMULAE

To obtain more accuracy one may take into account the second power of $(T_d - T_c)/T_c$ in the expansion of $\ln[1 + (T_d - T_c)/T_c]$ in (8) and in the same way as we got (7) one can get

$$T_c = [T_d\{\epsilon L/C_pT_d - 2 + \sqrt{(\epsilon L/C_pT_d - 1)^2 + 2\ln(T/T_d)}\}]/[2\epsilon L/C_pT_d - 3 + 2\ln(T/T_d)]. \quad (10)$$

There is another way to get more accurate formulae by taking into account only the first term in the expansion of $\ln[1 + (T_d - T_c)/T_c]$.

Since the difference between T_d and T_c is small, it is more convenient to write

$$T_c = T_{d1}[\epsilon L/C_pT_{d1} - 1]/[(\epsilon L/C_pT_{d1}) - 1 + \ln(T_d/T_{d1})]$$

where T_{d1} is the dew point temperature, a parcel of air would attain when lifted dry adiabatically from the initial level specified by dry bulb temperature T and dew point temperature T_d, to the level where its dry bulb temperature becomes T_d. It is easy to prove that

$$T_{d1} = (\epsilon L/C_p)/[\epsilon L/C_pT_d + \ln(T/T_d)]$$

By substituting T_d and T_{d1} for T and T_d in (7) one can reach to

$$T_c = (\epsilon L/C_p)(C-1)/C[C - 1 + \ln(1 + (C_pT_d/\epsilon L) \times \ln(T/T_d))] \quad (11)$$

where

$$C = (\epsilon L/C_pT_d) + \ln(T/T_d))$$

(11) is much more accurate than (7).

We can define T_{d2} as a function of T_d and T_{d1} in the same way as we have defined T_{d1} as a function of T and T_d. Then by substituting Td_1 and T_{d2} instead of T and T_d in (7) one can get more accurate formula than that of (11). So we can have more and more accurate formulae by defining T_{d3} or T_{d4} and so on and follow as we have proceeded from (7) to (11). The procedure followed from (7) to (11) can be repeated with (7)* to obtain the equation

$$T_c = E(D-1)/D[D-1+\ln(1+233.16A/E)] \qquad (12)$$

Corresponding to dew point depression which reaches 100°C in the meteorological ranges, the maximum difference between T_{d1} and T_{c2} is less than about 2°C. It follows that the maximum error given by (11) and (12) is less than about .002°C for the previously mentioned ranges.

In Table 1 some results of the computation of T_c by (7), (9), (10) and (11) are given in degrees centigrade. (7) appears twice once using L_{vw} and the other L_{vi}. The computation given in Table 1 is for all cases whether realistic or not.

Table 1. Computed values for T_c in degrees centigrade using different formulae

T°C	T_d°C	T_c°C given by				
		$(7)_1$	$(7)_2$	(9)	(10)	(11)
-60	-60	-60.00	-60.00	-57.63	-60.00	-60.00
-20	-60	-65.70	-64.95	-62.76	-65.68	-65.68
00	-60	-68.12	-67.07	-64.95	-68.09	-68.09
20	-60	-70.32	-69.00	-66.94	-70.28	-70.28
40	-60	-72.33	-70.77	-68.77	-72.28	-72.28

In $(7)_1$, (10) and (11) in Table 1, L_{vw} is used while in $(7)_2$ L_{vi} is used.

Table 2. Computed values for T_c in °C, cases where (9) can be correctly applied

T°C	T_d°C	T_c°C given by		
		$(7)_1$	$(7)_2$	(9)
-24	-38	-40.41	-40.09	-40.35
-14	-38	-42.02	-41.49	-41.75
-04	-38	-43.55	-42.82	-43.07

Columns (3) and (4) in Table 2 give the computed values of T_c for the same range of T and T_d in the table applying $(7)_1$ and $(7)_2$. This was done for the sake of comparison between these results and those obtained by applying (9). (9) is valid for such ranges of T and T_d in Table 2.

As expected, it is clear from Table 2 that values of T_c given by (9) always lie between those values given by $(7)_1$ and $(7)_2$. It is also noticed that $(7)_1$ and $(7)_2$ give underestimate and overestimate values of T_c as given by (9) respectively.

Using Teten's empirical formulae for the saturation vapour pressure over a plane surface, the same procedure can be followed, but it is not tried here.

4. SUMMARY AND CONCLUSION

When the saturation water vapour pressure is considered either over water or over ice, the temperature at the lifted condensation level T_c can be given either by (7) or (11). The mathematical expressions satisfying the above case are

either

$$T_d > -40°C$$

$$\ln(T/233.16) + (L_{vw}/C_p)(1/T_d - 1/233.16) < 0$$

or

$$T_d \leqslant -40°C$$

where -40°C is taken to be the temperature at which the saturation vapour pressure change from over water to over ice.

On the other hand, in the case when the saturation vapour pressure changes during the lift from over water to over ice either (9) or (12) can be applied to get T_c. The conditions for this case is

$$T_d > 40°C$$

$$\ln(T/233.16) + (L_{vw}/C_p)(1/T_d - 1/233.16) > 0$$

Within meteorological ranges and with dew point depression which reaches about 100°C we can apply either (7) or (9) which gives maximum error less than 0.05°C or (11) and (12) which give maximum error less than 0.002°C.

Procedures have been explained to have more and more accurate formulae whatever large the dew point is.

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APPENDIX

Suitable Expansion of $\ln T_c$

Among the possible expansions of $\ln T_c$ for approximate solution of (4) are the following:

$$\ln T_c = \ln T + \ln\left(1 + \frac{T_c - T}{T}\right) \quad (A1)$$

$$= \ln T - \ln\left(1 + \frac{T - T_c}{T_c}\right) \quad (A2)$$

$$= \ln T_d - \ln\left(1 + \frac{T_d - T_c}{T_c}\right) \quad (A3)$$

$$= \ln T_d + \ln\left(1 + \frac{T_c - T_d}{T_d}\right) \quad (A4)$$

$$= \ln \bar{T} + \ln\left(1 + \frac{T_c - \bar{T}}{\;}\right) \quad (A5)$$

$$= \ln \bar{T} - \ln\left(1 + \frac{\bar{T} - T_c}{T_c}\right) \quad (A6)$$

$$= \ln T_d + \ln\left(1 + \frac{T_c - T_d}{2\,T_d}\right) + \ln\left(1 + \frac{T_c - T_d}{T_c + T_d}\right) \quad (A7)$$

where $\bar{T} = (1/2)\,(T + T_d)$ and the temperatures are in degrees Kelvin.

Formula (A1), (A4) and (A5) when applied in (4) give second order equation in T_c at least while (A2) and (A3) can give equation of the first order. (A3) is the most suitable of all and is applied in this paper.

REFERENCES

Barnes, S.L., 1968: An empirical shortcut to the calculation of the temperature and pressure at the lifted condensation level. J. Appl. Met., 7, 511-511.

Haurwitz, B., 1941: Dynamic Meteorology, New York and London, McGraw Hill Book Co., Inc., 39-41.

Hess, S.L., 1959: Introduction to Theoretical Meteorology, New York, Holt, Rinehart and Winston, 362pp.

Homboe, J., G.E. Forsythe and W. Gustin, 1948: Dynamic Meteorology. New York, John Willey and Sone, Inc., London, Chapman and Hall Limited, 363pp.

Huschke, R.E., 1959: (Editor) Glossary of Meteorology, Boston Massachusetts, American Meteorological Society, 638pp.

Inman, R.L., 1969: Computation of temperature at the lifted condensation level. J. Appl. Met., 8, 155-158.

Lasheen, A.M., 1973: Approximations in the calculation of some convective elements. Progress Report MIRT (under preparation).

Petterssen, S., 1956: Weather analysis and Forecasting. Vol. 2 New York, McGraw Hill 40-42.

Prosser, N.E., and D.S. Foster, 1966: Upper air sounding analysis by use of an electronic computer. J. Appl. Met., 5, 296-300.

Saito, N., 1968: On the large scale structure of a disturbance in Easterlies over the Caribbean Sea. Tech. Memorandum ERLTM-NHRL 81, 8-9.

Stackpole, J.D., 1967: Numerical analysis of atmospheric sounding. J. Appl. Met., 6, 464-467.

ON THE CHANGES IN THE STRUCTURE OF WESTERLIES OVER EURASIA AND THE ONSET OF THE INDIAN SOUTHWEST MONSOON

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1. INTRODUCTION

India is essentially an agricultural country and its economy is very closely linked up with monsoon rainfall. The Indian southwest monsoon accounts for 70% of the annual rainfall over the Indian sub-continent. Most of the agricultural operations commence with the onset of the monsoon and an advanced knowledge of the date of onset of the monsoon over India will help to plan the agricultural operations efficiently.

The life cycle of the southwest monsoon, its onset in June, its advance over the sub-continent, its break periods, and final withdrawal in September-October, forms an important part of the Indian meteorology. Numerous workers have studied the various characteristics of the Indian monsoon, and it has been established that the onset and withdrawal of the southwest monsoon are part of the general circulation changes occurring in the tropical and sub-tropical zones (Sutcliffe and Bannon, 1954 : Yeh Dao and Li, 1959 : Lockwood, 1963 and Wright, 1967). Some attempts have also been made to find out relationships that exist between the onset of the monsoon and changes in the circumpolar westerlies of middle and high latitudes (Yin, 1949 : Flohn, 1960 : Ramaswamy, 1965 : de la Mothe and Wright, 1969 : Ramanadham and Visweswara Rao, 1970). But some of these studies were fragmentary in nature and could not categorically establish a close relation between the surge of the monsoon current and circulation to the north of the Indian region.

The present paper deals with the criterion used for the fixing of the dates of onset of the southwest monsoon over the Indian sub-continent, and with the large-scale changes in circumpolar westerly circulation in the northern latitudes over Eurasia that are associated with the onset of the monsoon.

2. FIXATION OF DATES OF ONSET OF THE MONSOON

The onset of the monsoon is not a single event but a series of successive events occurring within a fairly long period of time. Its arrival is a gradual process starting with the transition from extreme heat to a more humid atmosphere.

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There are different opinions about what really constitutes the onset of the monsoon. Most of the previous studies (India Meteorological Department, 1943; Bhullar, 1952; Ramdas et al, 1954) are all based on rainfall data. Identifying the onset of the monsoon based on rainfall is restrictive and not satisfactory, especially in regions where the rainfall caused by the pre-monsoon thunderstorm activity is comparable with the monsoon rainfall. This criterion leads to confusion because an airmass provides precipitation if certain dynamical processes are in operation in addition to the presence of water vapour. If these dynamical processes are absent or less active, the same airmass cannot provide precipitation. So it can be said that precipitation is an end product of a series of dynamical and thermodynamical processes associated with an airmass. Thus the precipitation criterion is not unique and it leads to doubtful identification of the onset of the monsoon current.

The monsoon current, it is well-known, is of maritime origin and the humidity characteristic which, though not completely conservative, does retain its identity after travelling over long distances. This maritime air can be easily distinguished from that of the continental air, so the authors felt that identification of airmass by the humidity mixing ratio to be more reliable than the precipitation from the same airmass.

The humidity mixing ratio data for 50 stations covering India are utilised in calculating five-day moving averages and plotted day by day from 1st May to 15th July for the years 1963, 1965 and 1966. The five day moving averages are taken to eliminate the local fluctuations. An interesting feature is at about the date of arrival of the monsoon current, the humidity mixing ratio for a particular five day average exhibited a sudden and well-marked rise over its two or three preceeding averages. Moreover the increase is not merely transient but is persistent characteristic of the monsoon airmass. The actual dates of arrival of the monsoon at individual stations for the year 1963 is presented in Fig.1. It is clear from the figure that the monsoon current arrived over the Malabar coast on June 3, 1963. Similar analysis showed that the monsoon arrived over this region on June 3 in 1965 and June 4 in 1966.

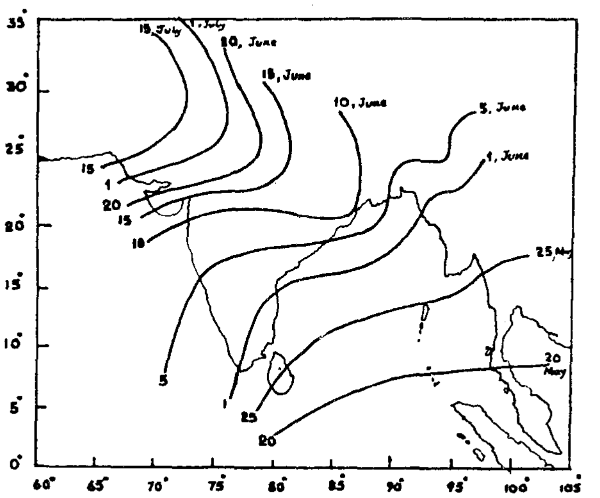

Fig.1. Dates of arrival of the monsoon current in 1963.

3. MIDDLE LATITUDE WAVE PATTERN OVER EURASIA

It has been recognised that the onset of the monsooon is part of general circulation changes occurring in the tropical and sub-tropical zones. These general circulation changes are associated with a reorganization of the wave pattern in the middle latitude westerlies. In order to determine the long-wave pattern, the 500mb surface contour charts were prepared from the data contained in the micro-films of the Northern Hemispheric Data Tabulations and analysed for the periods May 25 - June 5, 1963; May 21 - June 5, 1965; and May 25 - June 8, 1966. The heights of the 500mb surface are taken at the intersection points along a longitude and at the parallels of 40°, 45°, 50°, 55° and 60°N. The mean of these five values is the average height of the 500mb surface along a particular longitude and within the latitudinal belt of 40°-60°N. Such average height values are computed for every ten degrees of longitude between 20°W and 160°E. These values are plotted in a graph where the longitude and date form the X and Y axes respectively. Such diagrams represent detailed pictures of the long-wave profiles along the zonal belt. Fig.2 shows the long-wave profile for May 25 - June 8, 1966. Long-wave profiles along the zonal belt for the years 1963 and 1965, not presented here, show good similarity with the profiles presented here.

The long-wave pattern over Eurasia, during the transition period from pre-monsoon to monsoon period, shows a dominant one wave pattern with the ridge lying over central Asia near about 50°E. This ridge is flanked by two extended troughs lying over Europe and eastern Asia respectively. The wave length between these two troughs varied from 90° to 140° of longitude in the cases studied. Such large wave lengths more than 100° of longitude are observed whenever the height profiles over central and eastern Asia are flat. In such a situation, the eastern Asiatic trough lies in the western Pacific near 150° - 160°E. The presence of the Asiatic ridge near 50°E results in a marked northerly component both over and north of India.

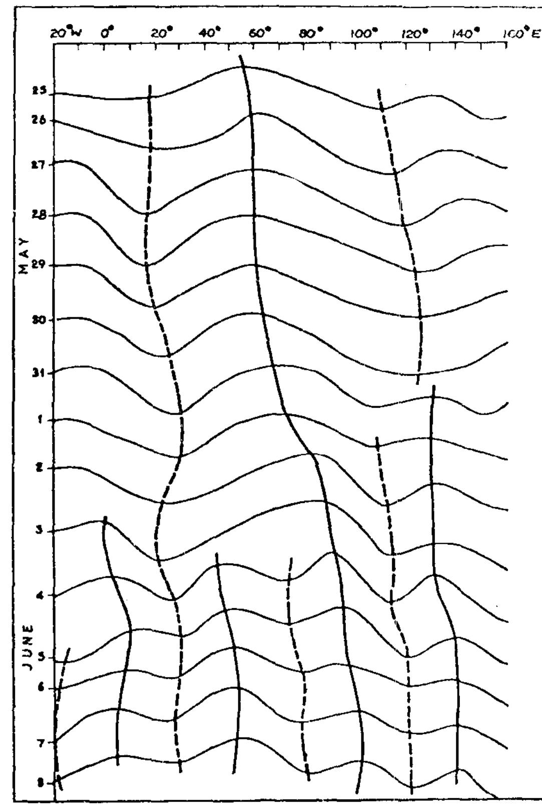

Fig.2. The average height of 500mb surface between 40° and 60°N extending from 20°W to 160°E for the period May 25 - June 8, 1966. Vertical dashed lines represent positions of troughs and vertical solid lines ridges.

The wave pattern described above shows a well-marked transformation at the time of one or two days before the onset of the monsoon over southern peninsular India. The long-wave pattern is featured with three waves over Eurasia with an average wave length of about 45° of longitude. Such a situation is due to the formation of new subsidiary troughs and ridges. The formation of such troughs is most likely to occur in regions which are directly or indirectly related to topographic factors as well as to thermal and frictional influences associated with land and water surfaces. Two new troughs along 20°W and 70°E are formed in the middle latitude westerlies while the trough near 120°E is sharpened. The date of formation and establishment of the new trough near 70°E is coincident with the date of onset of the monsoon over the southwest coast of India. Similar observations are reported by Yin (1949), Flohn (1960) and de la Mothe and Wright (1969).

4. LOW LATITUDE WAVE PROFILES

It is clear from Fig.1 that the monsoon air arrives over Burma earlier compared to India. In order to explain such a phenomenon, the low latitude wave profiles during the transition period are examined.

The height profiles are computed following the procedure mentioned earlier. The average height of the 500mb surface between 15° and 35°N at 5° intervals along a longitude are calculated. The average height values are computed at 10° intervals between 40° and 120°E. Such height profiles are prepared for the periods May 25 - June 8, 1963; May 21 - June 5, 1965 and May 25 - June 8, 1966. Fig.3 represents the wave pattern of 1966.

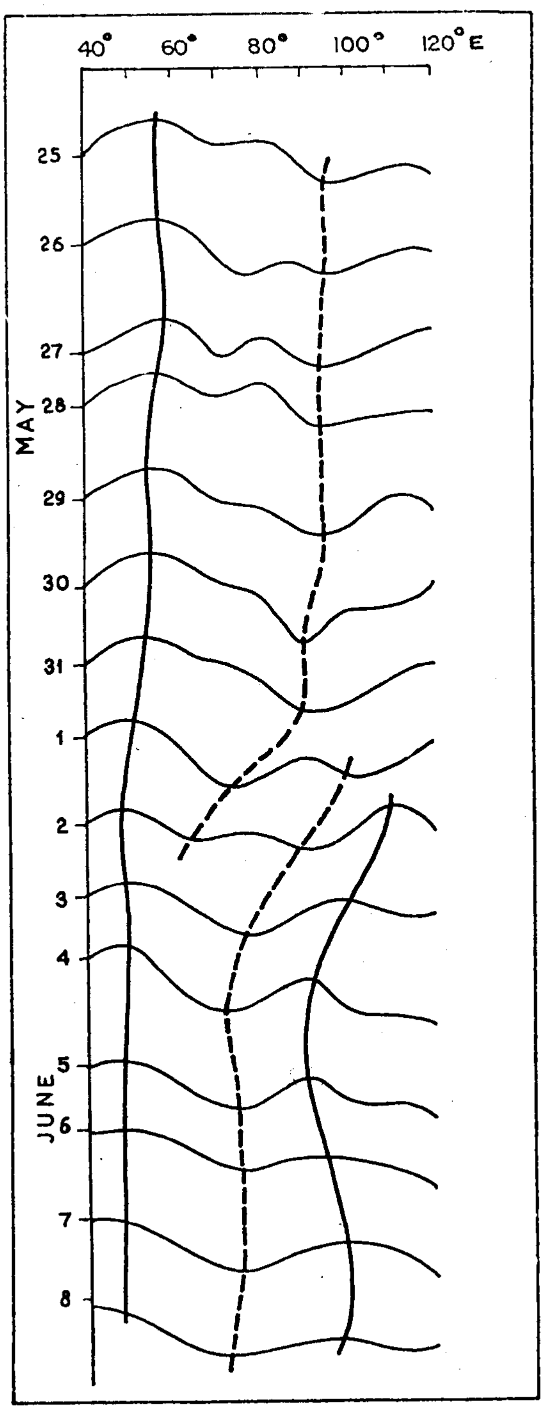

Fig.3. Same as Fig.2 for May 25 - June 8, 1966 between 15° and 30°N extending from 40°E to 120°E.

During the transition period an upper trough is lying between 90° and 100°E with a ridge to its west near 50° - 60°E. As a result there is a northerly component of the flow over India. This wave pattern showed breakdown two or three days prior to the onset of the monsoon over Malabar coast. The trough has split into two, with one branch retreating westwards and dissipating presumably due to the absence of middle latitude trough along 70°E; the second one which was at the original longitude also shifted westwards to near 70° - 80°E and became quasi-stationary about the date of onset of the monsoon. Such a shift has earlier been reported by Wagner (1931) and Yin (1949). Consequently the northerly component of the flow aloft over India is suddenly replaced by a southerly component at the same time as the sub-continent comes under the influence of the convergence east of the trough that must be present in order to effect the westward progression of the trough.

This has clearly brought out that Burma is always situated to the east of the upper-air trough, whereas India lies to its west during the pre-monsoon period. Superposition of the high tropospheric flow pattern and its attendant pressure field on the low-level circulation have the effect of accelerating the monsoon over Burma and of retarding it over India. After the trough shifts to 75°E, southerly wind components at high levels prevail over the entire Burma-India region and reinforces the monsoon everywhere.

5. STRUCTURE OF ZONAL WINDS AT 200MB LEVEL

Zonal winds along 75°E at 200mb averaged for three day intervals are examined during May and June of 1963, 1965 and 1966 in order to study the changes in structure of zonal winds. Fig.4 illustrates the zonal winds in 1966. The circumpolar westerlies extend to 15°N during the pre-monsoon period. The sub-tropical ridge line is also lying along this latitude. This ridge shifts rapidly northwards a few days before the onset of the monsoon. Consequently there is a rapid decrease of wind speeds over northern India and a considerable portion of the westerly stream is forced to retreat northwards. At the same time the region of west wind maximum which lies near 30°N in the pre-monsoon season shifts northwards to 38°N a few days prior to the onset of the monsoon.

Easterlies are observed to the south of the ridge line and these easterlies spread northwards and strengthen as the ridge line moves northwards. Consequently the easterlies over south India strengthen as the westerlies over north India weaken. These easterlies attain jet intensities in the upper troposphere at the time of the onset.

A secondary belt of strong westerlies is evident between 50° and 70°N. The intensity of the northern maximum is invariably less than the southern one. This secondary belt of wind maximum is probably due to the polar front jet stream, and it does not show any consistent variation of location because of its proximity to the polar region; hence its variation in position and strength may not be significant while considering the monsoon circulation.

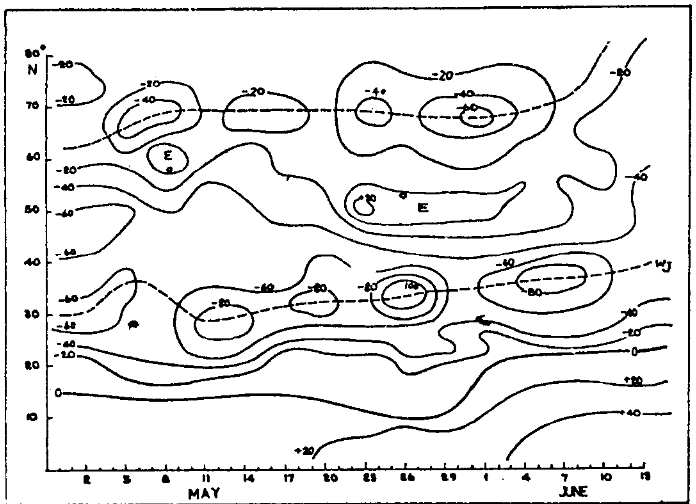

Fig. 4

Fig.4. Mean zonal winds at 200mb along 75°E averaged for 3 day intervals for the period May 1 - June 15, 1966. Isolines are labelled in knots with positions of maxima (jets) indicated by dashed line. Positive values represent easterlies.

Similar observations are reported by many earlier workers (Yin, 1949; Sutcliffe and Bannon, 1954; Ananthakrishnan and Ramakrishnan, 1965 and 1966; Wright, 1967).

6. DISCUSSION

Since the beginning of what might be termed the aerological era, roughly 1945, it has been recognised that there are large-scale interrelations between the flow aloft of high and low latitudes of the northern hemisphere and that these relations profoundly affect the tropical weather (Cressman, 1948a). In order to correlate the burst of the Indian monsoon with the circulation changes over Eurasia, the authors examined various aspects of the circulation.

Rapid polar circulation at the 500mb level is characterised by offshoots stretching out towards the equator. The jet stream and the polar front which lie at the equatorward side of the polar zone bulge out and reach lower latitudes. On the other hand weak polar circulation results in jet stream and polar front passing to higher latitudes. During the northern summer, the circulation in the northern hemisphere slows down, polar air retreats and the equatorial air expands towards the north. Simultaneously in the southern winter hemisphere the circulation intensifies resulting in extension of the westerly air-flow towards the equator. The equatorial air situated between these two convergence zones is then pushed back towards the warmer northern hemisphere. As the polar regions in the southern hemisphere are much colder than the polar region of the northern hemisphere, the winter circulation in the southern polar regions is much more vigorous than the circulation in the northern hemisphere in winter. This difference in the circulation pattern seems to be responsible for the deep push of the airmass from the southern hemisphere into the northern hemisphere during northern summer in the monsoon region.

Consequently a portion of the westerly stream and the associated sub-tropical jet stream that encircles the Himalayas along their southern border during the pre-monsoon period is displaced to the north of the mountains. Such deformation in the flow imposed by these huge mountain ranges causes a major readjustment of the low-latitude wave pattern aloft. As a result the low-latitude upper-air trough gets displaced rapidly from one steady position near 95°E to another relatively steady position near 75°E. In this position southerly components of the flow prevail aloft over India due to the convergence to the east of the trough line. Such a situation superimposed on the pressure gradient resulting from the large-scale differential heating results in the advance of the inter-tropical convergence zone.

During winter the Asian continent north of 40°N is intensely cold and the presence of Tibetan Plateau extends this cold area even farther south. As the sun moves north of the equator from about the beginning of April the snow line begins to recede rapidly and the Asian continent starts warming quickly and the thermal contrast weakens. Further, the cold air-flow weakens since the pole is less cold and receives insolation continuously and there is also a rapid degeneration of this cold air as it flows over the warm Asian continent. As a result the circum-polar westerly circulation weakens and is accompanied with the redistribution of the zonal current over Eurasia. Such a redistribution and weakening of westerlies are observed at the time of onset of the monsoon. Rossby (1939) and Cressman (1948b) demonstrated that strength of the zonal current is related to the motion and wavelength of the long-waves. After the transition from strong to weak circulation readjustment of wavelengths is necessary. Sometimes this readjustment takes place by way of relative motion of the troughs but frequently a new trough forms somewhere between the two extended troughs (Namias and Clapp, 1944). The formation of such new troughs is likely to occur in regions where topographic factors favour trough development.

Consequently the wave pattern consisting of one wave during the pre-monsoon period changes to three waves with an average wavelength of about 45° of longitude. One of the wave troughs of this new wave pattern becomes established near 70°-75°E. This is partly orographically imposed due to the northward displacement of the sub-tropical westerly jet stream and partly due to the decrease of wave lengths as earlier mentioned. The date of establishment of this new trough near 70°-75°E coincides with the date of onset of the monsoon over southwest coast of India (Yin, 1949; Flohn, 1960; de la Mothe and Wright, 1969).

Summarising the above it can be stated that the rapid reorganization of the middle latitude wave pattern coincides with the equally rapid phase shift of the low-latitude wave pattern, with the northward shift of the sub-tropical westerly jet stream and with the transition from pre-monsoon to monsoon regime over India.

The sub-tropical ridge line in the upper troposphere shifts northwards to 20°N and heralds the onset of the monsoon. In association with this the sub-tropical jet stream is induced to move northwards and a major portion of the westerly stream is detoured to the north of the high mountain ranges. The burst of the monsoon occurs after the core of the westerly jet stream shifts to the north of the Indian plains and probably beyond the Tibetan Plateau.

7. SUMMARY

The authors, in the present investigation, utilised the change in the humidity mixing ratio as an indicator of the arrival of the monsoon current. This seems to offer a better criterion for the arrival of the monsoon current than the rainfall which is being used by the Indian Meteorological Department.

The structure and intensity of westerlies in the middle and lower latitudes undergo drastic change at the time of onset of the monsoon. The sub-tropical jet stream and a portion of the westerly current over north India are induced to flow north of the Himalayas as a consequence of weakening of the northern polar circulation and the northward shift of the sub-tropical anticyclone in the upper troposphere. Such a redistribution and weakening of the westerly current are responsible for the transformation of the wave pattern into three wave structures in the middle latitude westerlies over Eurasia. Further the northward progression of the jet stream imposes the formation of the trough along 70°E in the middle latitudes due to the orography of the Tibetan Plateau. In addition, the northward displacement of the westerly current over north India is responsible for the rapid westward displacement of the low-latitude mean trough from about 95°E to near 75°E.

Thus, the broad-scale rearrangement of the long-wave pattern coupled with the shift of the sub-tropical westerly jet stream and the phase-shift of the low-latitude wave pattern seem to govern the advance of the monsoon over southwest peninsular India.

R E F E R E N C E S

Ananthakrishnan, R. and Ramakrishnan, A.R. 1965: Proc. Symp. Meteor. Res., IIOE, Bombay, pp 415.

Ananthakrishnan, R. and Ramakrishnan, A.R. 1966: Presented at the Fourth Tech. Conference on Hurricanes and Tropical Meteorology, Miami, U.S.A. (Unpublished).

Bhullar, G.S. 1952: Ind. J. Meteor. Geophys., 3, pp 25.

Cressman, G.P. 1948a: Dept. Meteor. Univ. Chicago, Misc. Rep. No. 24, pp 68.

Cressman, G.P. 1948b: J. Meteor., 5, pp 44.

de la Mothe, P.D. and Wright, P.B. 1969: Meteor. Mag., 98, pp 145.

Flohn, H. 1960: Symp. Monsoons of the World, New Delhi, pp 75.

India Meteorological Department, 1943: Climatological Atlas for Airmen.

Lockwood, J.G. 1963: Meteor., Mag., pp 92, pp 75.

Namias, J. and Clapp, P.F. 1944: J. Meteor., 1, pp 57.

Ramada, L.A., Jagannathan, P. and Gopala Rao, S. 1954: Ind. J. Meteor. Geophys., 5, pp 305.

Ramaswamy, C. 1965: Symp. Meteor. Res., IIOE, Bombay, pp 317.

Ramanadham, R. and Visweswara Rao, P. 1970: Symp. on Droughts, Andhra Univ., Waltair, India.

Rossby, C.G. 1939: J. Mar. Res., 2, pp 38.

Sutcliffe, R.C. and Bannon, J.K. 1954: Sci. Proc. Int. Assoc. Meteor., Rome, pp 322.

Wagner, A. 1931: Gerlands. Beitr. Geophysik, 30, pp 196.

Wright, P.B. 1967: Meteor. Mag., 96, pp 302.

Yeh, T.-C., Dao, S.-Y and Li, M.-T. 1959: The Atmosphere and sea in motion, Oxford Univ. Press, New York, pp 249.

Yin, M.T. 1949: J. Meteor., 6, pp 393.

ANALYSE DE DEUX PERTURBATIONS AYANT EVOLUE EN DEPRESSIONS TROPICALES AU LARGE DE DAKAR

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1 - INTRODUCTION

Les perturbations mobiles de l'Afrique Occidentale et Centrale semblent, dans plus de la moitié des cas, être à l'origine des dépressions ou cyclones tropicaux qui intéressent la mer des Caraîbes. Neil FRANK (4) trouve que 56 % de ces perturbations ont leur origine en Afrique. C'est ce qu'il appelle des "Dakar Systems". En particulier, en 1970, 11 "Dakar Systems" ont poursuivi leur déplacement jusque sur le Pacifique.

L'un des objectifs de G.A.T.E. sera l'analyse de ces perturbations.

ZIPSER (5) propose deux schémas de formation et d'évolution de ces perturbations nettement différents selon qu'il s'agit de lignes de grains ou de dépressions tropicales.

1.1 Cyclogénèse tropicale

- Une perturbation synoptique déclenche de la convergence dans les basses couches
- la convection s'organise et s'intensifie
- la zone de convection reste intimement liée à la perturbation (pas de downdrafts)
- net réchauffement du système par dégagement de chaleur latente au sein de la perturbation
- réalimentation et renforcement de la perturbation origine.

1.2 Lignes de grains

Même schéma général, sauf que :

- la zone de convection engendre des courants subsidents (downdrafts) qui affaiblissent la convection dans la perturbation mais la renforcent à l'avant
- la chaleur latente se dégage alors à l'extérieur de la perturbation
- destruction de la perturbation origine.

1.3 Conditions particulières à l'Afrique de l'Ouest

Les perturbations évoluant sur le continent africain différent de celles observées sur l'Atlantique Ouest ou le Pacifique :

- le flux qui les baigne n'est pas homogène
- la convergence interne de la mousson favorise la convection
- le Tropical Easterly Jet et l'African Easterly Jet favorisent souvent le développement de cellules convectives.

Les deux perturbations étudiées ci-après ont eu la même évolution, malgré leurs différences de structures à l'origine.

Elles ont été analysées élément par élément en essayant de montrer les points communs et les différences.

2 - ANALYSE DES DONNEES A.P.T.

2.1 1970

- 31.7.70 : Importantes masses nuageuses sur la Guinée et le Sud Sénégal
- 01.8.70 : Importante masse nuageuse sur le Sud Sénégal, accompagnée de pluies. D'autre part, ligne de grains sur la Boucle du Niger. A ce stade de l'analyse, on ne décèle pas l'onde d'Est existant sur le Mali.
- 02.8.70 : Circulation cyclonique sur le Sénégal
- 03.8.70 : Dépression tropicale centrée par 16°N/23°W.
- 04.8.70 : La dépression tropicale est maintenant centrée par 18°N/24°W.
- 05.8.70 : La dépression tropicale en voie de désagrégation, est centrée par 18°N/38°W.

Cette dépression a été répertoriée par FRANK (3) comme onde d'Est le 9 Août sur les Barbades.