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## A SEMI-QUANTITATIVE METHOD OF FORECASTING SUMMER STRATUS IN NORTH AMERICAN TROPICAL MARITIME AIR

By J. J. GEORGE and W. M. BRADLEY

[Eastern Air Lines, Hapeville, Ga., December 1936]

Of all North American air masses, probably the least variance in properties is observed in Tropical Gulf and Tropical Atlantic air (1). About the only marked variation noticed is the elevation at which the so-called Superior (2) air is encountered; and even this variation is of little or no significance for the subject under discussion. Because of this uniformity, these air masses (hereafter referred to collectively as Tropical Maritime air, since they are practically indistinguishable) should prove particularly amenable to a quantitative method of forecasting the attendant stratus deck.

During the summer, Tropical Maritime (TM) air is markedly unstable and is characterized by daytime cumulus, showers, and thunderstorms within the air mass. This is typical of a "Cold" air mass. "Cold" is used here in the same sense as in Bergeron's classification, i. e., the air is colder than the surface over which it is moving; it is therefore warmed from below, and instability results. Petterssen (3) and Vernon (4) have explained the California stratus as a result of such instability. This explanation does not, however, hold for the stratus formed in TM air because ordinarily no pronounced inversion is present at low levels to limit convection. During the night the lower layers of the air are cooled somewhat by radiation, and some stabilization takes place, as is shown by the cessation of the thunderstorms and by the normally clear nights.

Radiation or ground fogs of any intensity are practically unknown in this air mass in summer. The large moisture content (about 18 gr/kg at the surface) effectively limits radiation; and the diurnal temperature range in TM air is consequently very small, rarely amounting to more than 15° or 18° F. On the other hand, stratus fogs are common to the air mass, occurring frequently and covering large areas. The area visited by these fogs includes all the inland areas reached by the air mass itself, most commonly south of the 35th parallel and east of the 100th meridian. The actual formation of the fog may be, and often is, extremely rapid. Airline pilots flying in the area state that at times, when flying in clear air with the ground perfectly visible, the fog may form solidly below them during the time necessary to make a notation on the ship's papers. Although this type of stratus is important to airline operation, it is more important to the itinerant flyer not so familiar with local navigation aids and conditions, for the reason that almost invariably a ceiling is left under the fog layer, seldom if ever decreasing to less than 300 feet (sufficient for experienced pilots at most airports in the affected area); in addition, visibility is always good under the fog, so that, in general, aviation difficulties are only moderate.

Unfortunately, records of airplane soundings for this area were not readily available for previous years, and during the summer of 1936 a minimum of this type of stratus was observed east of the Mississippi River. For this reason the soundings presented here are limited to one station, San Antonio, Tex., and are for only the summer of 1936. They are, however, typical of the few soundings from Shreveport, La., and Montgomery, Ala. Even with limited aerological material, it is easy enough to obtain a fairly representative picture of the vertical distribution of temperature and moisture for various conditions of fog formation. Figure 1 shows four typical soundings represented on an ordinary temperature-elevation chart. The lapse rates in the turbulence layers all lie between 5° and 6° C per km, a value greater than the moist adiabatic for the high temperatures involved. The moisture distribution (not shown) is typical of a turbulence layer; that is, the specific humidity is nearly constant at a high value up to the base of the inversion, where it decreases slightly. The relative humidity is high throughout the layer but reaches a maximum at the base of the inversion. Table 1 compares the inversion base in these ascents with the pressure gradients and stratus formations.

TABLE 1

| Curve | Height above ground of inversion base | Geostrophic wind from isobars of 2000 chart preceding night | Geostrophic wind from following 0800 chart | Cloud deck |
|---|---|---|---|---|
| | *Meters* | *Force Beaufort* | *Force Beaufort* | |
| I | Zero | Near zero | Near zero | Clear. |
| II | 200 | 5.6 | 5.3 | 0.6 ST between 0600 and 0800 hours. Elevation unknown. |
| III | 600 | 8.0 | 5.3 | Broken stratus base 250; top 610 meters. |
| IV | 1,180 | 7.5 | 6.7 | Solid stratus base; top 1,000 meters |

Since ground fog of any intensity is almost unknown in this air mass, it follows that radiation alone is too ineffective to produce, or even to be a decisive factor in, the formation of the stratus deck. For this reason, attention is logically turned to the only factor present in the stratus formation which is absent when conditions are ideal for ground fog formation—a turbulence-produced inversion.

The soundings in figure 1 show the small magnitude of the inversion, and, to some extent, the relation between the elevation of the base of the inversion and the pressure gradient. It should be noted that for calm conditions the inversion is a surface one of small magnitude; as the velocity increases, the base of the inversion is forced upward, but the inversion produced is of the order of only 1° C. This small value is to be expected, considering the originally rather steep lapse rate in this air mass.

Having arrived at the tentative conclusion that turbulence is the necessary factor in the formation of the stratus

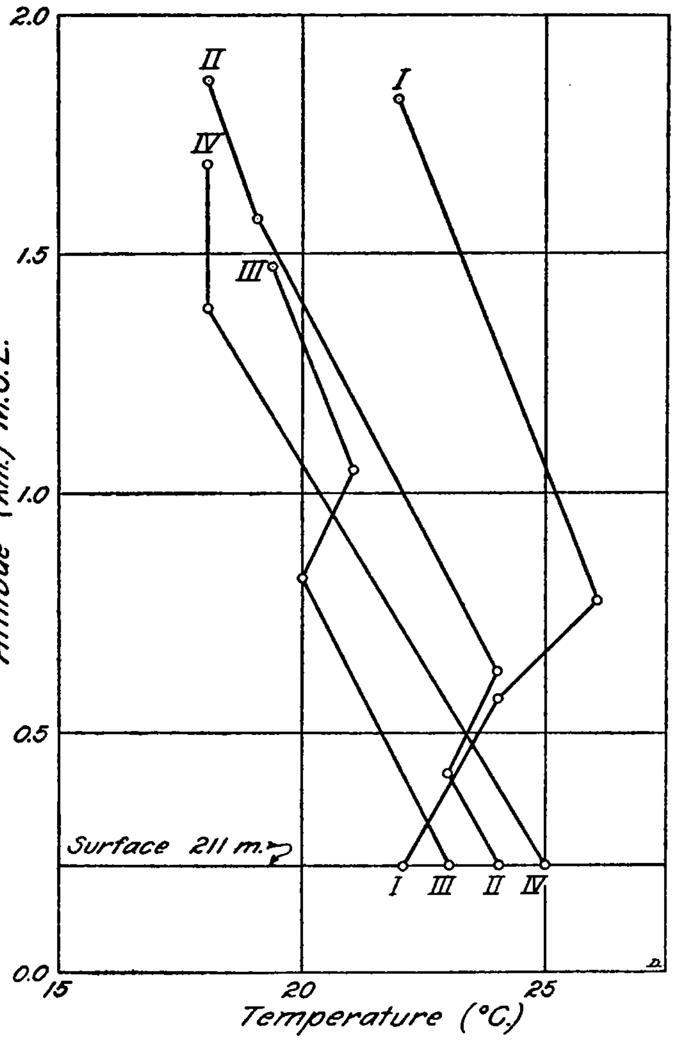


FIGURE 1.

and having verified the existence of such a turbulence layer from airplane soundings, we still must prove this conclusion and ascertain whether the stratus layer can be forecast by means of the condition that causes it. Rossby and Montgomery (5) indicate that for a given roughness parameter, the height of the layer of frictional influence is directly proportional to the wind speed at anemometer level. Although it should be possible to proceed along these lines, using the wind at anemometer level, a great deal of research at each individual station would be necessary to determine the indiviudal roughness factor. In addition, anemometer readings are not made with the precision which would be essential to use in these computations. Hence it was not considered practical to proceed on the basis of anemometer data.

Some measure of wind velocity however, is obviously the logical method of attack upon this problem of the formation of a turbulence inversion. The forecasters at the airway forecast center at Atlanta have long recognized that on nights when the stratus appeared the evening balloon run showed a definite maximum of velocity between 1,000 and 3,000 feet above the surface. This fact, of course, conforms to the well known Ekman spiral, and provides a close approximation to the gradient wind (**6**). Accordingly, a dot graph (fig. 2) was prepared by plotting wind direction against velocity at approximate gradient level as determined from the evening balloon runs at the Atlanta, Ga., airport station. The circles indicate nights on which no fog was observed; crosses, nights on which stratus formed; and diamonds, cases in which the stratus never became solid overcast. No case of formation of a stratus condition was noted for wind directions between east and west through north, partly because mari-

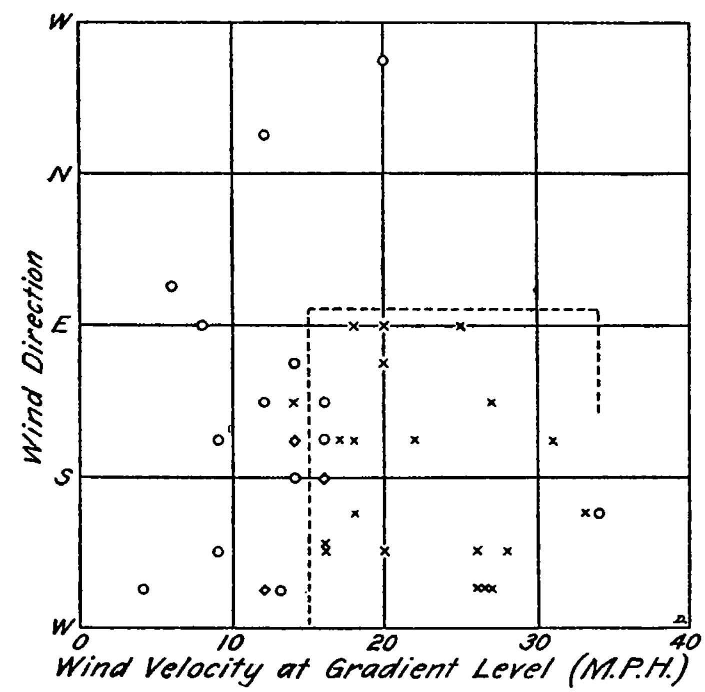


FIGURE 2.

time tropical air seldom if ever travels southward against the normal monsoon with sufficient velocity to establish the necessary turbulence layer, and partly because of slight foehn effect at Atlanta of winds from the northwest quadrant.

An amazingly sharp division between nights with, and nights without, the fog is immediately evident at a critical velocity of 15 m.p.h.; out of a total of 37 cases in which Tm air was present, only 3 are not thus segregated, and in each of these cases the wind velocity was not more than 1 m.p.h. from the critical velocity. A minimum velocity of about 15 m.p.h. must be attained before a turbulence layer, and consequently a stratus condition, can be formed. One case of no fog formation that occurred with a velocity of 34 m.p.h. will be covered in later paragraphs.

If, instead of plotting wind velocity against direction, we plot wind velocity against time of fog formation, we obtain figure 3. A smooth curve somewhat resembling an exponential curve is suggested, and has been drawn in the figure asymptotic to the critical velocity of 15 m. p. h. It should be noted that the time of actual formation of the stratus deck was in no case farther away from the curve than 2½ hours, and that the average is about 40 minutes. Such a relation is far too close to be coincidence; it seems justifiable to conclude that we have here ample confirmation of the original hypothesis, and at the same time a simple, accurate and direct method of forecasting the occurrence of the fog. This curve also explains the lack of fog formation for a velocity of 34 m. p. h. mentioned previously—the corresponding time of formation read from the figure is 0715, which in summer is so late that the sun has already destroyed the inversion produced by turbulence.

At velocities below 15 m. p. h., the turbulent layer is so shallow as to be negligible. The earliest times of fog formation should be expected with the lowest wind velocities, because mixing at the inversion surface is small and also because complete mixing in the homogeneous layer is

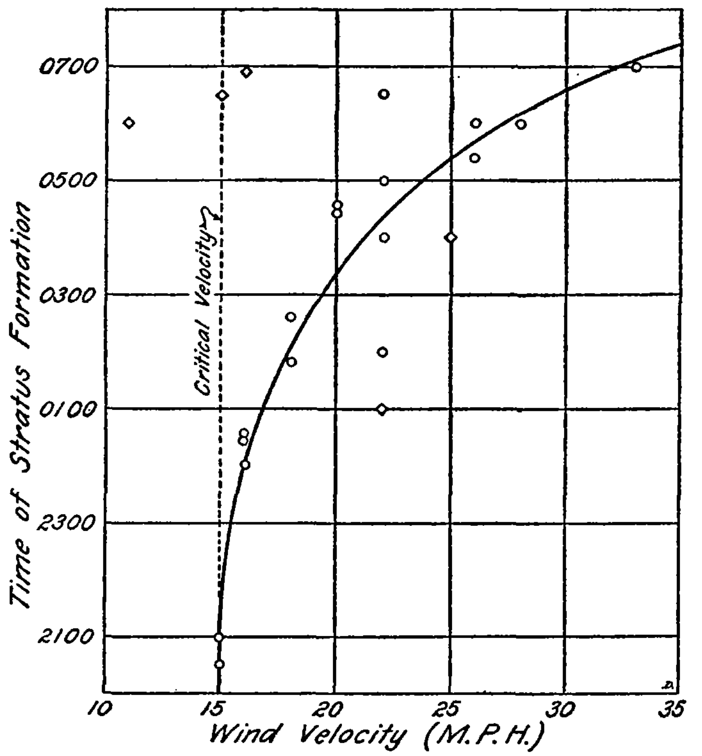


FIGURE 3.

more rapidly attained for the shallower layers. This statement is true under the logical assumption that the formation of the turbulence layer can begin only near sunset. That it cannot begin earlier is apparent from the unstable character of the air, which has no inversion to limit convection caused by either heating or turbulence. As velocities increase, turbulence extends higher, and a correspondingly longer time is required to form the thicker homogeneous layer. Nocturnal radiation of course assists in the formation of the stratus layer, since the latter forms only at night, but such assistance is comparatively minor because the extremely high moisture content limits the radiation; certainly its effect is quite uniform because of the homogeneity of the air mass.

A number of other interesting facts connected with the formation became apparent when the data were examined closely. It is true (fig. 3) that between the hours of 8:30 p. m. and 12:30 a. m., the curve has too great a slope for accurate determination of the time of fog formation; but this fact becomes of minor importance when it is considered that fogs which form early (before 2300 C. S. T.) are very likely to remain broken until well after midnight. This, too, is logical, since turbulence has not then had time in some places to form a pronounced inversion, while it has in other places, because of a slightly different character of terrain perhaps. Fogs which form late (after 0500) are also likely to remain broken, because of mixing by the attendant high wind velocities, heating of ground, and austausch.

Figure 4 shows an apparent relation between wind velocity and time of breaking of the stratus deck, for different approximate elevations of the base of the inversion. Unfortunately, the data are not sufficiently numerous to portray the actual relation. There should be very little relation between these two elements at the higher inversion levels, because of the indeterminate austausch taking place at the correspondingly high velocities, as well as other factors which cannot readily be measured; with only lower turbulent layers, mixing is slight and hence more regularity should be noted. The family of

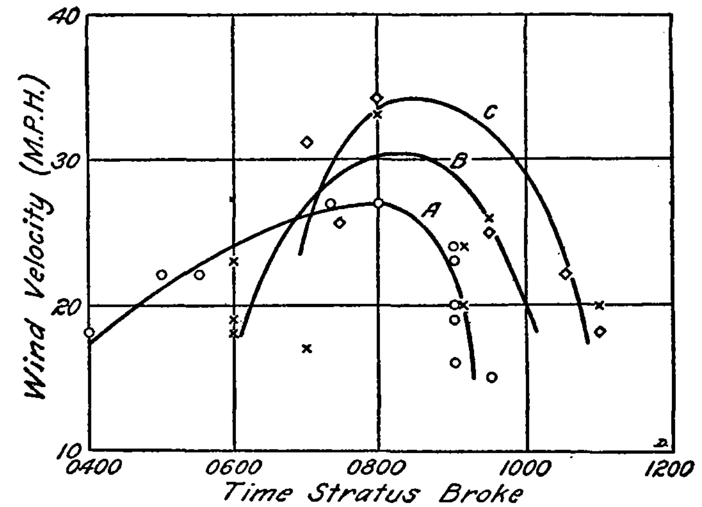


FIGURE 4.

curves in figure 4, marked A, B, and C, respectively, represent the relation for the approximate gradient levels of 1,500, 2,000, and 3,000 feet. Wind velocities are plotted as ordinates, and the times the stratus deck began to dissolve as abscissae. Circles represent an approximate elevation of the base of the inversion of 1,000 to 1,500 feet, crosses 1,500 to 2,500 feet, and diamonds 2,500 to 3,500 feet. For the lower levels—i. e., 1,000 to 1,500 feet—the curve fits the data so well that there can be little doubt that some such relation exists. For the higher levels, however, curves B and C were drawn from curve A as a pattern as much as from the actual data. Theoretically, we should expect only one time of breaking to be associated with one velocity. The fact that these curves seem to indicate definitely two possible times does not necessarily mean the theory is wrong. Rather, the few observations on the ascending branch of the curve indicate abnormal conditions of formation where the gradient wind is in the process of either increasing or decreasing. Even considering the numerous sources of inaccuracy here, it is probable that with more data and more airplane observations these curves would be verified.

It sometimes happens that balloon runs are not available for areas in which forecasts are desired. In view of this, some seventy analyzed 2000 E. S. T. charts, for the summers of 1935 and 1936, which contained Tm warm sectors were examined. The isobars for these sectors were redrawn with extreme care (all available airway reports were used, making a rather close network of reporting stations). A geostrophic wind scale constructed from the well-known approximate formula, $G=\frac{\Delta P/\Delta S}{\rho.2\omega\sin\phi}$, where G is the geostrophic wind, $\frac{\Delta P}{\Delta S}$ the pressure gradient, $2\omega\sin\phi$ the apparent deflective force of the earth's rotation, and $\rho$ the density of air. By applying the scale to the isobars, the theoretical geostrophic wind velocity may be read off directly in force Beaufort. A thorough check of these charts showed almost as definitely a critical velocity as the pilot balloon data. The velocity by this method was found to be half way in force 5 Beaufort, about 21 m. p. h., a value slightly higher than that determined by the balloon runs. It is probable that the explanation for the higher value is partly the method of determining the gradient wind from the balloon runs: If precision methods were employed, values about 2 m. p. h. higher would be obtained. It is true that the geostrophic wind scale would give too low values of gradient winds because of the usual slightly anticyclonic curvature of the isobars. It may be that at the higher elevation (about 300 m) of Atlanta, not as great a wind velocity is required to produce the stratus (because of additional cooling by lifting) as in the remainder of the area to the west where the wind scale was principally employed.

Little difficulty was encountered in forecasting, from the isobars, the areas over which the fog would form, except in about 10 of the seventy cases; in each of these cases, the pressure gradient changed during the night; a sufficient gradient was present on the 8 p. m. chart, but had diminished to below the minimum on the 8 a. m. chart, and as a result either no fog was formed or else it was much more broken than was indicated. A close check of these exceptions proved that it was a comparatively simple matter to draw the isallobars and apply Petterssen's kinematical methods (7) to the movement of isobars; since the extreme horizontal homogeneity in Tm air (1) demands parallel isobars, and since we are only interested in their relative movement, the computations are so simplified that they may in most cases be done by inspection. The appropriate formula is $h=h_0+(C_2-C_1)t$, where $h_0$ is the original distance between isobars, $h$ is the distance after a time $t$, and C refers to the instantaneous velocity of isobars 1 and 2. The velocity of an isobar is $C_i=-Th$, where T is the three hour pressure tendency. With sufficient accuracy for our purposes, the unit distances may, in most cases, be considered equal for the neighboring isobars; this assumption permits a simple subtraction of pressure tendencies at the two isobars, which is then multiplied by 4 in order to obtain the spreading of the isobars between the two 12 hour charts.

It is interesting to note that in no case was there observed a condition where the gradient was insufficient on the 8 p. m. map to produce the stratus and had strengthened sufficiently during the night enough to produce it. The probable explanation lies in the time required for the actual height of the homogeneous layer to attain the theoretical height. Rossby and Montgomery state that such time is of short duration; nevertheless, mixing at the bounding surface is going on constantly and is, no doubt, sufficient to prevent the formation of the stratus.

REFERENCES

(1) H. C. Willett, American Air Mass Properties, *M. I. T. Meteorological Papers*, vol. II, no. 2.
(2) H. C. Willett, Discussion and Illustration of Problems Suggested by the Analysis of Atmospheric Cross Sections, *M. I. T. Meteorological Papers* vol. IV, no. 2.
(3) Sverre Petterssen, On the Causes and the Forecasting of the California Fog, *Journal of the Aeronautical Sciences*, vol. 3, no. 9.
(4) E. M. Vernon, The Diurnal Variation in Ceiling Height Beneath Stratus Clouds, MONTHLY WEATHER REVIEW, January 1936.
(5) C. G. Rossby and R. B. Montgomery, The Layer of Frictional Influence in Wind and Ocean Currents. *M. I. T. Meteorological Papers*, vol. III, no. 3.
(6) F. Exner, *Dynamische Meteorologie*, p. 121.
(7) Sverre Petterssen, Kinematical and Dynamical Properties of the Field of Pressure with Application to Weather Forecasting, *Geofysiske Publikasjoner*, vol. X, no. 2.

# TEMPERATURE AND RAINFALL CHANGES IN THE UNITED STATES DURING THE PAST 40 YEARS

By LARRY F. PAGE

[Bureau of Agricultural Economics, Washington, D. C., December 1936]

About 3 years ago, Kincer (1) showed that the annual average temperature at a number of stations in the United States and elsewhere had been rising for a period of 20 to 30 years or more. For a few records, the four quarters of the year were separated, and the results indicated some differences in trends. A recent publication of the U. S. Geological Survey (2) includes an examination of precipitation trends at stations grouped in 15 sections of the United States, by three seasons, December to April, May to August, and September to November. Ten-year moving averages were used, and indicated upward trends in the fall, and declining trends in winter and summer precipitation [(2), p. 48]. The question naturally arises as to the similarity of trends among the months, within the seasons. Since the divisions of the year, as used in these studies, represent somewhat different types of weather influence, it might be reasonable to assume that the variations in trend were due to these differences. We might expect, at least, a shading from one average seasonal trend to the next for the same station or area.

To investigate this, State monthly average temperature and precipitation records were used, as published in *Climatological Data* by the U. S. Weather Bureau. In order to simplify calculation and comparison, only data for the years 1896 to 1935, inclusive, were used with the exception of California, where the published records begin with 1897. No indication is given, from this study, as to the future trend. Such changes as are shown might be due to fluctuations or periodicities of about 40 years *or longer*, or to more or less permanent changes in our climate. This 40-year period was divided into two series of 20 years each, 1896 to 1915 and 1916 to 1935. The mean value and standard deviation[1] for each of these series for each State were computed, and from the latter

[1] Standard deviations were computed by the formula $\sigma_{x_1}=\sqrt{\frac{(\Sigma x_1-\bar{x}_1)^2}{n_1}}$. Preference is often given to the formula using $n-1$ in place of $n$, but differences are small, and certain considerations indicated the use of the former. See (3), p. 51.