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MR. E. G. BILHAM, B.Sc., D.I.C.

Mr. E. G. Bilham, Deputy Director (Forecasting) retired from the Meteorological Office on May 30, 1953, after 38 years' service.

Mr. Bilham graduated with honours in Physics from Imperial College, London, in 1913, and joined the staff of the Meteorological Office in April 1915 as resident observer at Kew Observatory. While at Kew Observatory he analysed the observations of the level of water in a well for tidal and meteorological effects, and later published the results in a series of papers in the *Proceedings of the Royal Society* and the *Quarterly Journal of the Royal Meteorological Society*. He was transferred to the Benson aerological observatory in 1916, and from Benson to the Forecast Division at Headquarters, South Kensington, in 1917. He remained a forecaster for seven years, and in that period wrote papers on isallobaric analysis and on the synoptic meteorology of summer fog at Scilly.

In 1924 Mr. Bilham was promoted Superintendent and given charge of the Instruments Division. He was posted in 1930 to be Superintendent of the British Climatology Division which included the British Rainfall Organization, and held that post until the end of 1938. During his fourteen years in the Instruments and British Climatology Divisions he invented a height computer for aerological work and a humidity slide-rule; he also wrote many papers on a large number of subjects including methods of testing instruments, climatology, and the fall of hailstones. His study of the frequency of intense falls of rain, published in *British Rainfall 1935*, is a very important contribution to the application of rainfall data in engineering. At this time, too, he wrote the book which immediately became the standard work on its subject, "The climate of the British Isles", published by Macmillan in 1938.

Mr. Bilham returned to synoptic meteorology in January 1939 as Superintendent of the Headquarters Forecast and Civil Aviation Division. Soon afterwards he had to solve the complex problems associated with the evacuation of the Headquarters forecast organization to Birmingham in August 1939 and its transfer to Dunstable in February 1940. The war greatly increased the work and size of the Forecast Division and brought the introduction of new techniques such as "sferics" and upper air analysis. Mr. Bilham was promoted Assistant Director (Forecasting) in 1942, and on the re-organization in 1947 was promoted Deputy Chief Scientific Officer and appointed Deputy Director for Forecasting.

Mr. Bilham described the methods and organization of weather forecasting in "Here is the weather forecast", a book for the general reader published in 1947.

He has been a member of the Commission for Aeronautical Meteorology of the International Meteorological Organization and of the Commission for Synoptic Weather Information of the International Meteorological Organization and its successor, the World Meteorological Organization; in these capacities he has represented the Meteorological Office at several important international conferences.

Mr. Bilham's work has been characterized by ingenuity and practical skill in the improvement of meteorological instruments and techniques generally, and his ability for clear exposition has been of great value in the writing of such official publications as the "Observer's handbook" and "Your weather service". He may well claim to have left his imprint on all the major divisions, Instruments, Climatology and Forecasting, of which he has been in charge.

Mr. Bilham is an accomplished pianist and has always been ready to place his talent at the disposal of the staff at official parties. His retirement will be a considerable loss to the Meteorological Office which he leaves with the good wishes of all those with whom he has been associated.

The Director expressed the good wishes of the staff to Mr. Bilham at a meeting in Victory House on May 30 and presented him on their behalf with a camera and a cheque towards a motor lawn mower. Mr. Bilham in giving his thanks told interesting and amusing anecdotes of his early years at Kew and Benson and of some of those under whom he had served.

THE VARIATION OF WIND NEAR THE TROPOPAUSE

By E. GOLD, F.R.S.

The interesting result found by Austin and Bannon[1], that on occasions of strong upper winds, 70 kt. or more, the maximum wind occurs 30 to 40 mb. (3,000 to 4,000 ft.) below the level of the tropopause, means that the thermal wind is reversed before the tropopause is reached. In general the isotherms in the upper troposphere are approximately parallel to the isobars (or the contour lines) and temperature and pressure increase together in a horizontal direction. The rate of increase of temperature is of the order of 1°C./100 Km., though it may be sometimes much more, or much less, rapid. So long as this normal thermal structure is maintained the wind will increase with height. If there were no horizontal gradient of temperature the geostrophic wind would decrease slowly with height in the troposphere. From equation (1) below, the horizontal gradient of temperature necessary to counterbalance this decrease is

$$M = 2\omega V \sin\phi \frac{L}{g}$$

where the letters have the meanings given below. For a geostrophic wind of 90 kt. and a lapse rate of 8°C./Km. the value of M is 2×10^{-8} °C./cm. or 1°C./500 Km.

The change of wind as the tropopause is approached, or passed, is dependent on the thermal structure, and may give useful information confirming or correcting the recorded temperatures and the deductions based on them.

The major factor producing the change of wind (or *vice versa*) is the slope of the tropopause. As a rule, if there is no slope, i.e. if the tropopause is practically horizontal, there will be no marked variation in the character of the change of wind with height as the tropopause is approached or passed. This is not a common case and is unlikely to occur where the upper wind is of the order of 70 kt.; it is much more likely to be found in a region of light upper winds.

If the tropopause, as it generally does, slopes upwards towards the higher pressure and higher temperature (and therefore intersects horizontal planes along isobars or contour lines) the wind will decrease in passing through the tropopause. If there is an inversion, there will be a corresponding sudden decrease of wind, followed by a gradual decrease with height. If there is no inversion the decrease of wind will be gradual throughout. Unless there is a decrease in the lapse rate in the troposphere before the inversion is reached the decrease of wind will begin only at the inversion and not below it.

If the slope of the tropopause is upwards from the high pressure towards the low, the wind will increase up to and through the tropopause unless the horizontal gradient of temperature in the troposphere is reversed and is opposite to the gradient of pressure. In that case the wind would decrease up to the tropopause but would begin to increase at the tropopause or soon after.

If the lapse rate in the troposphere, after being constant or increasing, begins to decrease at a distance h below the tropopause and continues to decrease up to the tropopause, the wind will also begin to decrease before the tropopause is reached, and the maximum wind will, in this case and only in this case, occur below the level of the tropopause. It is emphasized that h is measured from the tropopause so that the height at which the decrease of lapse rate begins increases as the tropopause rises. The wind would not begin to decrease in the troposphere if the decrease of lapse rate began at a fixed height above sea level.

The amount of the change in the geostrophic wind for these different conditions can be expressed in terms of the lapse rates, the magnitude of inversion of temperature, the horizontal gradient of temperature, the distance h and the geostrophic wind itself.

Four general cases can be taken:—

(i) Lapse rate L up to the tropopause. Inversion of t°C. at the tropopause, i.e. in a relatively small distance. Lapse rate zero above the inversion.

(ii) Lapse rate L up to tropopause, l above tropopause. Discontinuity of lapse rate at tropopause.

(iii) Lapse rate L up to level at distance h below tropopause. Uniform decrease of lapse rate to zero at tropopause. Lapse rate l above tropopause.

(iv) Lapse rate in the troposphere increasing (or decreasing) towards the higher pressure.

In all cases the tropopause slopes at an angle A and the tropopause contours are parallel to the isobars or to the contours of the isobaric surface at tropopause level.

The following symbols are used; the values quoted are used in the examples:

g gravity acceleration. $9{\cdot}8 \times 10^2$ cm./sec.^{2}
R constant in equation $p = R\rho T$. $2{\cdot}87 \times 10^6$ cm.^{2}/sec.^{2}°C.
ω angular velocity of earth. $7{\cdot}3 \times 10^{-5}$ radians/sec.
ϕ latitude. For latitude 50°, cosec ϕ is $1{\cdot}3$
λ $2\omega \sin \phi$. For latitude 50°, λ is $11{\cdot}2 \times 10^{-5}$
T temperature in degrees Absolute. Tropopause temperature 220°A.
p pressure

L lapse rate. 8×10^{-5} °C./cm. (4·4°F./1,000 ft.)
l lapse rate. 2×10^{-5} °C./cm. and -2×10^{-5} °C./cm.
M horizontal gradient of temperature in the troposphere. 10^{-7} °C./cm. (1°F./30 miles approximately)
A angle of slope of tropopause. $\tan A = 1/100$
V geostrophic wind
v change in geostrophic wind
z vertical co-ordinate, positive upwards
x horizontal co-ordinate at right angles to isobars and positive towards increasing pressure.

The rate of change of geostrophic speed with height, under the conditions specified, is readily derived from the fundamental relations between pressure, temperature, density and geostrophic wind.

From the relations $$\frac{dp}{dx} = \lambda \rho V, \quad \frac{dp}{dz} = -g\rho$$

it follows, by differentiating the first with regard to z and the second with regard to x that

$$\lambda \rho \frac{dV}{dz} = -\lambda V \frac{d\rho}{dz} - g\frac{d\rho}{dx}.$$

By using the result $d\rho/\rho = dp/p - dT/T$ derived from $p = R\rho T$, the equation becomes

$$\lambda T \frac{dV}{dz} = \lambda V \frac{dT}{dz} + g\frac{dT}{dx} \qquad \ldots\ldots (1)$$
$$= \lambda VL + gM,$$

and the total change of speed in a height z is, approximately,

$$v = -\frac{V_m Lz}{T_m} + g\frac{Mz}{\lambda T_m} \qquad \ldots\ldots (2)$$

where T_m is the mean temperature and V_m the mean speed in the atmosphere from 0 to z.

At an inversion, sloping at an angle a, Lz is $-t$ and M is $\pm t/z \cot a$ the plus or the minus sign being taken according as the slope is upwards towards the high or the low pressure. Also $V_m = V$ and $T_m = T$, z = thickness of inversion layer and $t°$ the increase of temperature.

Thus at an inversion $$v = \frac{Vt}{T} \pm gt\frac{\tan a}{\lambda T}. \qquad \ldots\ldots (3)$$

These equations will now be applied to evaluate the changes in the cases mentioned; the first term of equation (2) is usually small, only 1 or 2 kt. for a change of height of 1,000 m., and the first term of equation (3) is usually smaller still.

Inversion $t°$ at tropopause.—From (3) with $a = A$, i.e. a slope of 1/100, the value of the second term in latitude 50° is $4t$ m./sec. or $8t$ kt. Slopes even steeper than 1/100 may occur, e.g. on January 5, 1950 there was a slope of 1/50 south of Ireland, but these very steep slopes extend for only 100 miles or so. When the slope is nearly uniform for 500 or 600 miles it is more likely to be 1/200, as for example on January 11, 1952, when the tropopause was about 20,000 ft. higher over the Mediterranean than over England with contour lines nearly equally spaced running approximately from west to east across France and Spain. But clearly, sudden decreases of 20 kt. or more may occur when there is a marked inversion at the tropopause. The value of the term Vt/T is only $\frac{1}{2}$ kt. if V is 110 kt. and t is 1 °C. and may therefore be disregarded.

If the slope were reversed, i.e. upwards towards the low pressure, the wind would increase $8t$ kt. at the tropopause with a 1/100 slope.

In the stratosphere the horizontal gradient of temperature is $M - L \tan A$ and as $L \tan A$ is 8×10^{-7} while M is only 10^{-7}, the geostrophic wind decreases very rapidly in the stratosphere above a steeply sloping tropopause. At a height z the decrease is $gz\,(L \tan A - M)/T$. With the values quoted above this gives a decrease of 28 m./sec. or 55 kt. at a height of 1,000 m. above the tropopause.

If the slope were 1/200 the decrease at the tropopause would be halved, $4t$ kt., but the decrease in the stratosphere would be only 3/7 as great, i.e. 12 m./sec. or 24 kt. at 1,000 m.

Lapse rate changing at the tropopause from L to l.—There is no sudden change in V at the tropopause in this case. The decrease above the tropopause would be very slow except for the horizontal gradient of temperature which is $M - (L - l) \tan A$. Thus the total change in a height z is by equation (2)

$$v = -\frac{V_m l z}{T_m} + \frac{gz(M - (L - l)\tan A)}{\lambda T_m}.$$

For the values quoted and $z = 1{,}000$ m. the decrease, disregarding the small first term, is 20 m./sec. or 39 kt. If l is negative, i.e. increasing temperature in the stratosphere, the decrease is 36 m./sec. or 71 kt.

If the slope were 1/200 the decreases would be 16 and 31 kt. instead of 39 and 71 kt.

If M were large and the slope small $(L - l) \tan A$ could be less than M and the wind would increase in the stratosphere.

Lapse rate decreasing to zero at the tropopause from the value L at and below a level distant h beneath the tropopause.—If z is measured from the level h below the tropopause, the horizontal gradient of temperature, in this layer of decreasing lapse rate, is $M - L \tan A . z/h$ and the lapse rate is $L\,(1 - z/h)$. Thus in this layer the integration of equation (1) leads to the following equation for the change of geostrophic speed,

$$v = \frac{V_m L z}{T_m}\left(1 - \frac{z}{2h}\right) + \frac{gz}{\lambda T_m}\left(M - L \tan A \frac{z}{2h}\right).$$

The first term is, in general, small compared with the second. The wind begins to decrease near the level

$$z = \frac{2Mh}{L \tan A} = \frac{h}{4}$$

if L is 8×10^{-5} and M is 10^{-7}.

If h is 1,000 m. the decrease up to the tropopause is 12 m./sec. or 24 kt. Above the tropopause the horizontal gradient of temperature is $M - (L - l) \tan A$ and the further decrease of wind is:

$$v = \frac{V_m l z}{T_m} + \frac{gz}{\lambda T_m} (M - (L - l) \tan A).$$

At a height of 1,000 m. in the stratosphere, this additional decrease is 39, 55 or 71 kt. according as l is 2×10^{-5}, zero, or -2×10^{-5}.

Lapse rate in the troposphere increasing (or decreasing) towards the higher pressure.—The variation of lapse rate may be represented by $L \pm Bx$ and the horizontal gradient of temperature is then $M \mp Bz$. Therefore the change of wind is

$$v = -\frac{V_m z}{T_m} (L \pm Bx) + \frac{gz}{\lambda T_m} (M \mp \tfrac{1}{2}Bz).$$

For a change in lapse rate of 1°C./Km. in a horizontal distance of 100 Km., which would be a rather rapid change, the value of B is 10^{-12} and $\frac{1}{2}Bz$ is only equal to M when z is 2,000 m. ($M = 10^{-7}$). If M were zero the change in speed would be 4 kt. in 1,000 m. and 16 kt. in 2,000 m. It appears likely therefore that a horizontal variation of lapse rate is only a subsidiary cause of decrease of wind before the tropopause is reached and that the main cause is a decrease of lapse rate commencing about 1 Km. below the tropopause. There will, however, be many occasions when the lapse rate is changing both horizontally and vertically in a sense tending to decrease the wind, but the magnitude of the changes is not sufficient to outweigh the increase due to M but only to reduce the rate of this increase, in the layers just below the tropopause.

REFERENCE

1. AUSTIN, E. E. and BANNON, J. K.; Relation of the height of the maximum wind to the level of the tropopause on occasions of strong wind. *Met. Mag., London,* **81,** 1952, p. 321.

GLOBAL CIRCULATION OF AIR AT HIGH LEVELS AND MECHANISM OF CHANGE OF TROPOPAUSE LEVEL

By A. H. R. GOLDIE, D.Sc.

In earlier papers the writer has referred to the effects on tropopause level and stratospheric temperature of the vertical component of circulation, both as between high and low pressure systems in specific cases[1] and as between tropical and temperate latitudes in the longer-period average global circulation[2].

It is many years since W. H. Dines[3] pointed out that the mean temperature of the atmosphere at the equator differs much less from the mean in England than it would do if temperature were solely a matter of radiative equilibrium, and that the factor making for equalization is, of course, circulation.

The chief purpose of this paper is to discuss the hypothesis that the air which constitutes the lower stratosphere of middle and high latitudes comes continually from the upper troposphere of the tropics and returns eventually to the troposphere of the polar regions in winter. This is one side of the global circulation. The other side in which air of the lower troposphere of middle and high latitudes drifts equatorwards, is heated and rises, is not discussed in this paper.

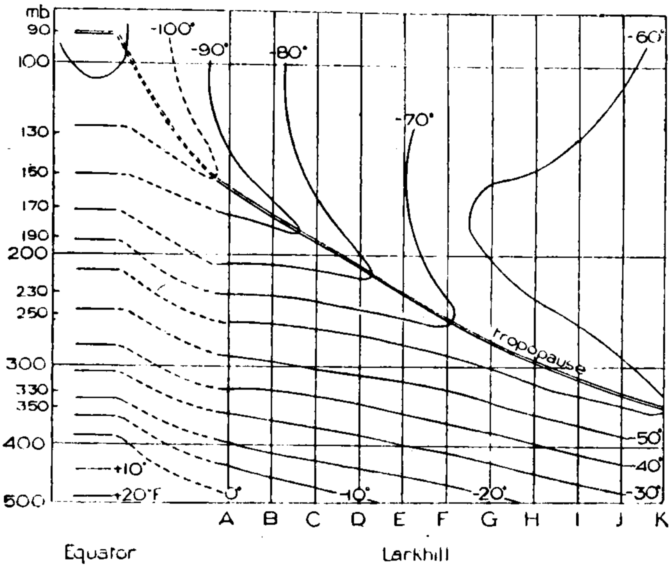
The recent classification by Bannon[4] of temperatures in the upper troposphere and lower stratosphere in England according to tropopause pressure provides a starting point, and also prompts some further discussion of the local patterns of air flow above and below the tropopause which might lead to the variety of temperature distributions found in temperate latitudes.

Typical distributions of temperature.—The data of Fig. I of Bannon's paper[4] have been used to plot the right-hand side of the upper part of Fig. I of the present paper. The figures used for column A are mean temperatures at the various pressure levels when tropopause pressure lies between 159 and 140 mb., those in column B relate to tropopause pressure between 179 and 160 mb. and so on. Isopleths of temperature have been drawn in, and also the position of the tropopause. The result may be taken as giving a typical average cross-section from high tropopause to low tropopause in temperate latitudes. On the extreme left of the picture are given mean values for the equator. The two sets of isopleths are connected by dotted lines.

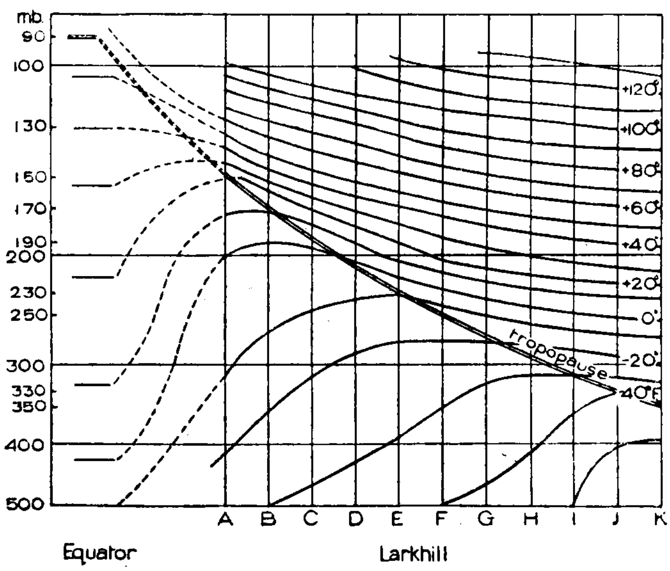
The lower part of Fig. 1 is a picture of the same data transformed into potential temperatures for a pressure of 400 mb. The conditions from A to K in these two diagrams are all to be found typically at one time or another in the month of January (or indeed generally in winter) in the southern part of the British Isles. Fig. 2 represents similarly the typical distributions for July. By what processes of movement and/or radiation can such variety of temperature distributions be brought about in a given month of the year?

Evidence for an average global meridional circulation.—It is obvious from all these figures that the tropopause is an important discontinuity or boundary in the pattern of temperature. From the characteristic wind distribution across the tropopause it is known similarly that there is in general a discontinuity in air movement. Typically the wind increases with height up to a maximum a short distance below the tropopause. There is then a very steep drop in wind speed on passing into the (relatively) slow-moving air of the stratosphere.

There is another feature which establishes not only a discontinuity of conditions between troposphere and stratosphere but a discontinuity of recent past history of the air and which points to an almost complete lack of mixing across the boundary. This feature is the customarily very low frost point in the lower stratosphere, a frost point for which there is no obvious explanation other than that the stratospheric air of temperate latitudes has in its recent past history come through a temperature equivalent to that in the upper troposphere of the tropics. Brewer[5] has discussed certain aspects of this subject. The coincidence of the range of frost points found in the lower stratosphere over England[6] with the range of temperatures found in the upper troposphere over the equator is too remarkable to be acceptably explained on any other ground at present conceived. Table I shows the percentage frequency of the frost points in the tropopause and in the stratosphere respectively for levels of observations from 350 mb. to 187 mb. Not only are the frost points, level for level, much lower in the stratosphere than in the troposphere, but they are grouped more closely. It is found also that in the troposphere the higher the temperature, the higher, on the average, the frost point. In the stratosphere the frost point is either quite independent of temperature or possibly on the whole falls as temperature rises. If values at the extremes are excluded as being possibly unreliable, it is found that about 90 per cent. of the stratospheric frost points lie within $-79°$ to $-111°$F. (or $211°$ to $193°$A.). The higher and lower values here correspond to the average temperatures in the tropics at 13 Km. and at 17 Km. (the tropopause) respectively. The low frost points found are thus consistent with drift of air from the upper 4 Km. of the tropical troposphere into the lower stratosphere over England.



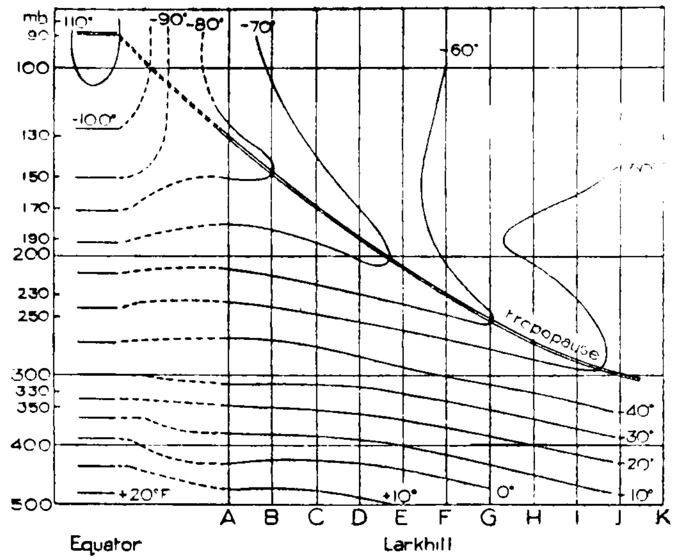
Actual temperature



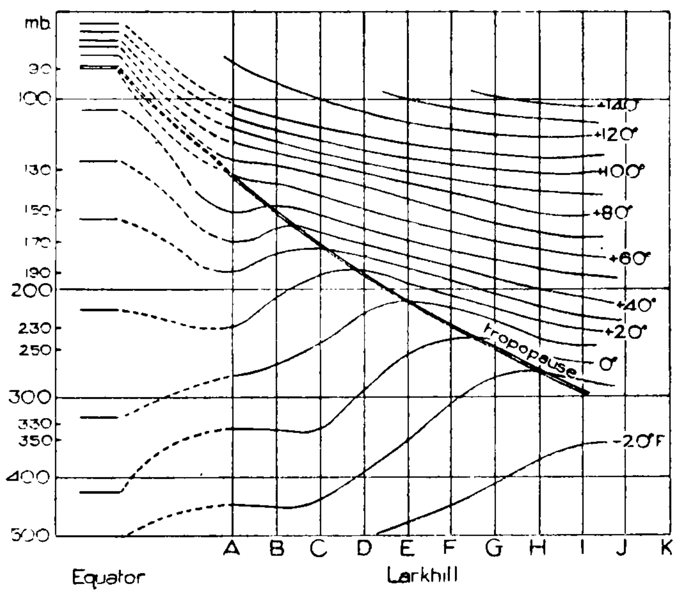
Potential temperature (400 mb.)

FIG. I—JANUARY TEMPERATURE AT LARKHILL IN RELATION TO TROPOPAUSE LEVEL AND AVERAGE JANUARY TEMPERATURE AT EQUATOR

There is, however, other (and much earlier) evidence, namely from the phenomena following the Krakatoa eruption[7] of August 27–28, 1883, for the actual existence of a drift of this sort. According to the discussion by E. Douglas Archibald, the evidence as to the spread of the optical phenomena, and in particular the evening glows, showed "that while in the equatorial zone the manner in which the first appearances succeeded one another, demands a due westerly current of considerable velocity, that in which they afterwards spread into higher latitudes in the northern hemisphere, shows a north easterly trend . . . the approach of the glow phenomena, both towards Europe and towards America, seems to have been from the south west".



Actual temperature



Potential temperature (400 mb.)

FIG. 2—JULY TEMPERATURE AT LARKHILL IN RELATION TO TROPOPAUSE LEVEL AND AVERAGE JULY TEMPERATURE AT EQUATOR

The tropical upper "westerly current" (i.e. E. wind) referred to was found to have a speed of about 76 m.p.h. It is known now from upper wind measurements that this at once identifies the region of fast travel with the upper troposphere within the tropics, say within some 13 to 17 or 18 Km. height, and possibly in some longitudes the lower stratosphere.

The average height of the principal part of the stratum producing the twilight glows was determined at various places in Europe and America as being about or slightly over 17 Km. It ranged from about 30 Km. in the tropics at first to about 13 Km. in high latitudes several months later. The twilight glows were most in evidence in middle latitudes from November 1883

TABLE I—PERCENTAGE FREQUENCY DISTRIBUTION OF FROST POINTS

Frost Point	Troposphere 350–187 mb. (609 cases)	Troposphere 250–187 mb. (163 cases)	Stratosphere 350–187 mb. (323 cases)
°F.		per cent.	
−33 or higher	1	0	0
−34 to −36	1	0	0
−37 to −39	1	0	0
−40 to −42	2	0	0
−43 to −45	4	0	0
−46 to −48	3	0	0
−49 to −51	3	1	0
−52 to −54	5	0	0
−55 to −57	7	1	0
−58 to −60	6	1	0
−61 to −63	7	4	1
−64 to −66	9	5	0
−67 to −69	9	6	1
−70 to −72	7	8	2
−73 to −75	7	7	3
−76 to −78	6	11	2
−79 to −81	6	13	7
−82 to −84	4	10	6
−85 to −87	4	10	7
−88 to −90	3	7	9
−91 to −93	1	4	10
−94 to −96	2	5	10
−97 to −99	1	4	10
−100 to −102	1	2	9
−103 to −105	0	1	8
−106 to −108	0	0	5
−109 to −111	0	0	6
−112 to −114	0	0	2
−115 to −117	0	0	1
−118 to −120	0	0	1

to February 1884, and had certainly reached latitude 60°N. by the end of November. This indicates a total time for the spread in latitude from Krakatoa (about 6°S.) as some 60 to 90 days to cover some 50° to 66° of latitude, the rate of the northward movement being slower as the latitude increased. This provides not only evidence that a drift of the sort postulated above does exist but also a measure of the time scale for air to get from the tropical upper troposphere and lower stratosphere to the stratosphere of temperate latitudes. The drift may not be steady, it may vary with longitude and season, but in some measure it must always go on.

The further discussion below will therefore be based on the feature that the air of the lower stratosphere of temperate latitudes has in fact in its recent life history got into the stratosphere via the upper troposphere of the tropics, and that the temperatures found over England in the stratosphere are the combined result of high-level tropical conditions, plus circulation including vertical motion, plus gain or loss of heat by radiation in the interval of time since leaving the tropics. Further comment on the causes of the lowering of the tropopause in middle and high latitudes is given on p. 206.

Dynamical effects accompanying the drift of air.—There are several points now to be noted about Figs. 1 and 2. The first is the relatively small gradient of potential temperature with height in the tropical troposphere, a condition consistent with global thermal circulation in which the heat is applied to the air principally in the tropics at the ground. Below the level shown on these diagrams the potential temperature in the tropics actually rises as the ground is approached.

The second point about Figs. 1 and 2 is that, level for level, both the actual and potential temperatures increase in the troposphere from temperate to low latitudes and in the stratosphere from low to temperate latitudes. The former is understandable enough in terms of the insolation, whether with or without a horizontal movement of air towards the equator. The latter (like the frost points found) can only continue to exist if northward drift with mass subsidence of air of higher potential temperature is continually in progress.

It is further to be noted, however, that the isopleths of potential temperature are not parallel to the tropopause but have only about half the slope of the tropopause. If there were no loss of heat on the journey from tropics to temperate latitudes and in the process of subsidence, the isopleths of potential temperature would be very widely spaced and would be parallel to the tropopause surface.

Without knowing the actual trajectories of the air masses (which may be devious) certain consequences can be seen. Compare air just above the tropopause in column A of Fig. 1 with air just above the tropopause in column H. If the stratospheric air all came into the temperate-latitude tropopause from the tropics, at or above the tropopause of column A, then the air just above the tropopause at H has subsided roughly 13,000 ft. more than that at A since leaving the tropics. With an actual temperature of $-98°$F. at 150 mb. at A, it ought in the course of such subsidence to have risen in temperature to $-22°$F. at 290 mb. Actually the temperature is only about $-66°$F., i.e. subsidence should have gained for it 76°F. and the actual gain is 32°F. The loss by radiation at H as compared with A is thus some 58 per cent. of the gross adiabatic gain in getting from A to H.

The amount of heat lost by air in getting from the tropics to temperate latitudes with a tropopause at, say, 150 mb. can be deduced, but in this case an estimate is first required as to the level at which the air—found eventually near the tropopause in temperate latitudes—leaves the tropics. This estimate can be got through the frost points. Table II, based on data from *Geophysical Memoirs* No. 88[6] shows the mean actual frost points in each season as measured by the Meteorological Research Flight over southern England at the tropopause and at levels 25, 50 and 75 mb. above the tropopause, and the mean actual pressure at these levels. Between these columns has been inserted the pressure at which, on the average, is found at the equator an air temperature equal to the mean actual frost point obtained by the Meteorological Research Flight. It is believed that in the upper troposphere of most of the tropical regions of the globe the air is normally saturated, i.e. the air temperature in the tropics may be taken as being also very approximately the frost point. Thus, if the stratospheric air as observed over southern England came originally from the tropical upper troposphere and has continually subsided, its frost point can be taken as a pointer to the level at which it left the tropics. The difference between the mean actual pressure and the corresponding equatorial pressure giving the amount of subsidence is included in Table II as well as the mean actual air temperature observed by the Meteorological Research Flight and the temperature

TABLE II—SEASONAL EFFECTS OF SUBSIDENCE AND RADIATION DURING MERIDIONAL MOTION ON TEMPERATURE OVER SOUTHERN ENGLAND

P_c = pressure at tropopause

Pressure level	Mean frost point	Corresponding equatorial pressure	Mean actual pressure	Amount of subsidence	Mean actual temperature	Temperature with adiabatic subsidence from equator	Gain or loss by radiation
	°F.	mb.	mb.	mb.	°F.	°F.	°F.
			WINTER (Jan. to Mar.)				
$P_c - 75$	−102	110	176	66	−58	−48	−10
$P_c - 50$	−98	120	201	81	−62	−41	−19
$P_c - 25$	−98	120	226	106	−68	−28	−40
P_c	−89	160	251	91	−72	−37	−35
			SPRING (Apr. to June)				
$P_c - 75$	−105·7	98	174	76	−53	−41	−12
$P_c - 50$	−101·4	112	199	87	−57	−37	−20
$P_c - 25$	−93	130	224	94	−64	−31	−33
P_c	−81·5	155	249	94	−67	−28	−39
			SUMMER (July to Sept.)				
$P_c - 75$	−104	105	164	59	−49	−53	+ 4
$P_c - 50$	−99	117	189	72	−52	−45	− 7
$P_c - 25$	−90·4	147	214	67	−57	−47	−10
P_c	−76·8	167	239	72	−63	−35	−28
			AUTUMN (Oct. to Dec.)				
$P_c - 75$	−105	100	215	115	−51	−17	−34
$P_c - 50$	−98	120	240	120	−52	−21	−31
$P_c - 25$	−90·6	147	265	118	−61	−23	−38
P_c	−79	164	290	126	−63	−11	−52

which the air observed by the Meteorological Research Flight would have had if the subsidence had been under adiabatic conditions. The difference between these last two columns gives the apparent gain (+) or loss (−) of temperature by radiation.

It will be seen that the four seasons have certain features in common. There is a considerable loss of heat by radiation at and just above the tropopause. At the tropopause indeed the average loss of temperature amounts to 72 per cent. of the gross adiabatic gain. At somewhat higher levels in the stratosphere, however, there is less loss and in summer there is even a small gain of heat by radiation. It must be pointed out, however, that whilst these calculations start from average conditions at the equator, the observations by the Meteorological Research Flight are somewhat selective. The temperatures at the higher levels happen to be a good deal warmer than the long-period seasonal means for Larkhill.

Theoretical extension to higher levels.—In Table III are summarized the mean amounts of subsidence taken from Table II. In the main, the amount of subsidence decreases with height. Each layer that drifts from equator to temperate latitudes is passing into ever narrowing circles of latitude and in addition to subsiding might be expected to be converging and increasing in thickness. If the convergence were proportional to the narrowing of the circles of latitude (which it need not be, because the relative speeds are also involved), then a thickness of about 45 mb. at the equator, or alternatively a thickness of 49 mb. at latitude 23°, would give 75 mb. in southern England.

From the corresponding equatorial pressure given in Table II it will be seen that the initial "equatorial" thicknesses (corresponding to 75 mb. in England) are 50 mb. in winter, 57 mb. in spring, 62 mb. in summer and 64 mb. in autumn.

TABLE III—SUBSIDENCE OF AIR FROM EQUATORIAL UPPER TROPOSPHERE TO LOWER STRATOSPHERE OVER SOUTHERN ENGLAND

P_c = pressure at tropopause

	WINTER (Jan. to Mar.)	SPRING (Apr. to June)	SUMMER (July to Sept.)	AUTUMN (Oct. to Dec.)
			millibars	
$P_c - 75$	66	76	59	115
$P_c - 50$	81	87	72	120
$P_c - 25$	106	94	67	118
P_c	91	94	72	126

It is important to get an idea of what happens at higher levels. For these there are as yet no frost-point measurements. One way of examining the problem is to extrapolate the mean amounts of subsidence from Table III on the assumption that the subsidence tails off to zero at zero pressure. On this basis the initial equatorial pressure of air arriving in England at the following pressure levels can be estimated as follows:—

Pressure level over England	Initial equatorial pressure of air Winter	Summer
mb.	mb.	mb.
75	38	45
100	52	63
125	67	78
150	82	95

Most of these levels imply starting points in the lower stratosphere of the tropics which is in accord with Goldie[2].

Table IV can now be compiled in which, however, the average temperatures over Larkhill are used rather than the temperatures observed by the Meteorological Research Flight from $P_c - 50$ mb. upwards. The result, for about a 170-mb. thickness above the tropopause, is seen to be that there is large radiation loss in the lowest part (as before), much smaller loss, especially in summer, about the 125-mb. level and large loss again at very high levels, especially in winter.

If the resultant gain or loss for the stratosphere up to the 75-mb. (18-Km.) level is averaged in the layers, weighted according to density, to 150 mb., a mean loss of 40°F. is obtained in winter, which is some 62 per cent. of the gross adiabatic gain, and a mean loss of 17°F. or 36 per cent. in summer. Probably the explanation of the differences with level is simply that where convection is inoperative radiation exchange within the system works continually towards the production of an isothermal condition. Too much significance should not, however, be attached to individual levels, because the individual figures could be altered a good deal by different assumptions as to the way in which subsidence changes with height. At low pressure a small change of pressure has, of course, a large adiabatic effect. It is to be noted that in England[8] in summer, a short way above the highest level (18 Km.) included in Table IV the temperature begins to rise steeply, and that at 30 Km. (or a pressure of about 13 mb.) the mean temperature is about −35°F.

TABLE IV—RADIATION EFFECTS ON TEMPERATURES OVER LARKHILL

P_c = pressure at tropopause

Pressure level	Equatorial temperature	Equatorial pressure	Mean pressure (Larkhill)	Amount of subsidence	Mean temperature (Larkhill)	Temperature with adiabatic subsidence from equator	Gain or loss by radiation
	°F.	mb.	mb.	mb.	°F.	°F.	°F.
			WINTER (Jan. to Mar.)				
$P_c - 175$	−86	38	75	37	−72	+ 3	−75
$P_c - 150$	−90	52	100	48	−69	− 7	−62
$P_c - 125$	−100	67	125	58	−69	−26	−43
$P_c - 100$	−110	82	150	68	−68	−42	−26
$P_c - 75$	−102	110	176	66	−70	−48	−22
$P_c - 50$	−98	120	201	81	−72	−41	−31
$P_c - 25$	−98	120	226	106	−68	−28	−40
P_c	−89	160	251	91	−72	−37	−35
			SUMMER (July to Sept.)				
$P_c - 175$	−89	45	75	30	−61	−24	−37
$P_c - 150$	−97	63	100	37	−63	−40	−23
$P_c - 125$	−108	78	125	47	−63	−54	− 9
$P_c - 100$	−106	95	150	55	−63	−52	−11
$P_c - 75$	−104	105	164	59	−64	−53	−11
$P_c - 50$	−99	117	189	72	−62	−45	−17
$P_c - 25$	−90	147	214	67	−57	−47	−10
P_c	−77	167	239	72	−63	−35	−28

Going back now to the time scale deduced previously, the result is that air moving from the tropical upper troposphere and subsiding to constitute the lower stratosphere of temperate latitudes has to lose in, say, an average of about 75 days (possibly more or less according to season) some 40°F. in winter and some 17°F. in summer out of the gross gain which it would have made under completely adiabatic conditions. This seems reasonable—half a degree Fahrenheit a day in winter and a quarter of a degree in summer.

The low tropopause of high latitudes.—At the thermal equator the tropopause is always high, close indeed to the height (17·5 Km.) computed by the author on the principle of global thermal circulation[2]. Even in high latitudes, the tropopause can be high if the pressure distribution is favourable to it; and in middle latitudes it can on occasions be very low. It might be said that the whole or most of the variety of conditions in middle and high latitudes arises from advection, and the author was probably the first to give weight to this effect[9]; but the upper wind observations accumulated since that time show that, though average latitude stratospheric characteristics may tend to be advected, the troposphere cannot simply carry the same stratospheric air with it. This does not mean that pressure distribution and circulation of hemispherical dimensions may not earlier have produced the features that were advected. Experience suggests that the tropopause cannot become very low, or remain very low, unless in a region surrounded by a strong cyclonic circulation. If such a circulation diminishes in intensity, the tropopause rises. The evidence indicates that the primary control on tropopause level in middle and high latitudes is tropospheric circulation, rather than latitude or temperature. If the troposphere becomes very cold in any area, the barometric gradient increases rapidly with height and favours a circulation around that area increasing with height up to the tropopause, and this is precisely the sort of circulation which should tend dynamically to lower the tropopause (see V. Bjerknes[10]) and warm the stratospheric air by subsidence.

Continued loss of heat from the stratosphere in the arctic winter may in itself tend to obliterate the tropopause, i.e. the diagram of Fig. 2 continued farther to the right would tend to a condition in which the potential temperature at the tropopause is no more than at the ground. It is most probably in this way that, in the arctic and antarctic, air is returned from stratosphere to troposphere. It can be shown that subsidence with radiation losses of the same order would suffice. Consider some examples of air starting with the potential temperature of the upper troposphere of the tropics and travelling to the arctic or antarctic with subsidence to various levels. In each case the starting condition will be taken as −115°F. at 90 mb. This air if it subsided to the 9, 6, 4 and 3 Km. levels with no heat lost in the process would rise to temperatures of +31°, +97°, +140° and +163°F. respectively. Actually the average temperatures found in midwinter at some of the coldest stations in high latitudes at these levels are about −61°, −49°, −25° and −20°F. From this it is seen that to arrive at the appropriate temperature the radiation losses accompanying the subsidence require to be 63, 69, 65, and 66 per cent. respectively. Conversely, radiation losses of slightly higher order would appear to render inevitable the exchange of stratospheric air near the poles of cold with tropospheric air of surrounding regions. Considered along its trajectory, the air may be gaining in actual temperature but it is continually losing potential temperature. The extreme temperatures in the upper air imply greater percentage losses and presumably involve loss of heat by radiation over longer periods in the polar winter. For example, the lowest found at Maudheim is given as −130°F. (−90°C.) at about 11 Km. In such a case the radiation loss would exceed the adiabatic gain by 30 per cent.

The more rapid changes of tropopause level.—What has been said previously related to the more gradual processes of tropopause change taking place as air from high levels in the tropics finds its way to positions in temperate latitudes. The changes in stratospheric and tropospheric temperature which might take place as superposed effects within an air mass in the shorter-period changes, such as the deepening or filling up of depressions, may now be considered. Since the scale of time is now in hours or days as against some 75 days, the extent to which radiation reduces the adiabatic gains or losses of temperature is presumably correspondingly less. It would thus be reduced to the order of one twentieth of the values estimated previously, i.e. the radiation loss would be of the order of 3 per cent. of the gross adiabatic temperature change. Where the loss is of this order, 3 per cent. (or even up to 10 per cent.) the difference in percentage loss at different levels should not seriously affect the conclusions reached below. Probably, in view of the actual temperature distribution, the maximum loss or gain of heat by radiation is close to the tropopause on either side.

Consider first curves D and F of Fig. 1: that is the mean temperature-pressure curves for Larkhill for January for tropopause pressures in the ranges 219–200 mb. (D) and 259–240 mb. (F) respectively.

Regarding these temperature distributions as derivable dynamically from one another, the amount of ascent or subsidence of air required at each level can be calculated to convert curve D, with mean tropopause 210 mb., to curve F, mean tropopause 250 mb. For this purpose, and as averages are being considered, the stratosphere has been taken for purposes of computation as being initially approximately isothermal and the troposphere as having the average lapse rate of 3·5°F./1,000 ft. Thus subsidence or ascent of 1,000 ft. implies in the stratosphere a gain or loss of temperature of 5·4°F. (the adiabatic rate), but in the upper troposphere of only 1·9°F. (the difference between the adiabatic and the average lapse rate). The result is shown in Fig. 3. Each

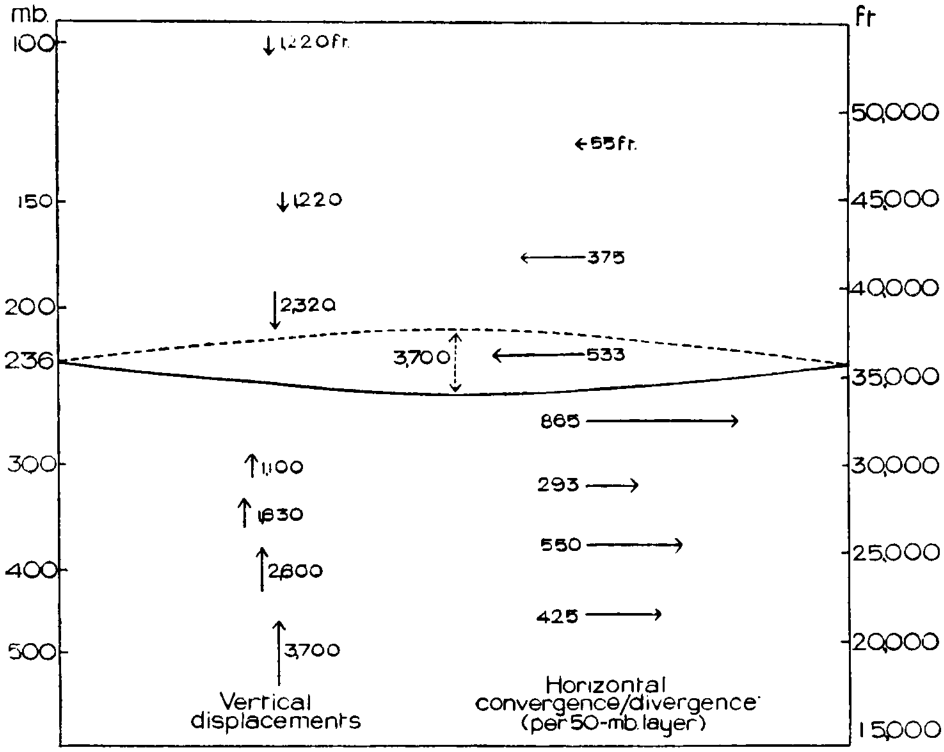


FIG. 3—CHANGE OF TROPOPAUSE FROM 210 MB. TO 250 MB. DURING JANUARY

vertical arrow indicates the required amount and sense of vertical displacement in feet, at the level at the centre of the arrow. The dotted arrow indicates the vertical movement of the tropopause. From the vertical displacements thus indicated it is possible to calculate the relative convergence or divergence required at different levels in the form of aggregate inward or outward displacement of 50-mb. slices of the atmosphere around the area within which the tropopause change is supposed to take place. These convergences and divergences and their sense are similarly indicated by horizontal arrows on a scale of their own.

Fig. 4 represents the results of a similar calculation applied to a rather extreme example taken from October. It has to be remembered that to include the effect of radiation, all the displacements shown in this figure would have to be increased by possibly 3 to 10 per cent.

The points of interest attaching to the mechanism thus shown are as follows:—

(i) Each case has maximum vertical displacement downwards in the stratosphere just above the tropopause. In the troposphere from the 500-mb. level up to near the tropopause there is considerable upward motion.

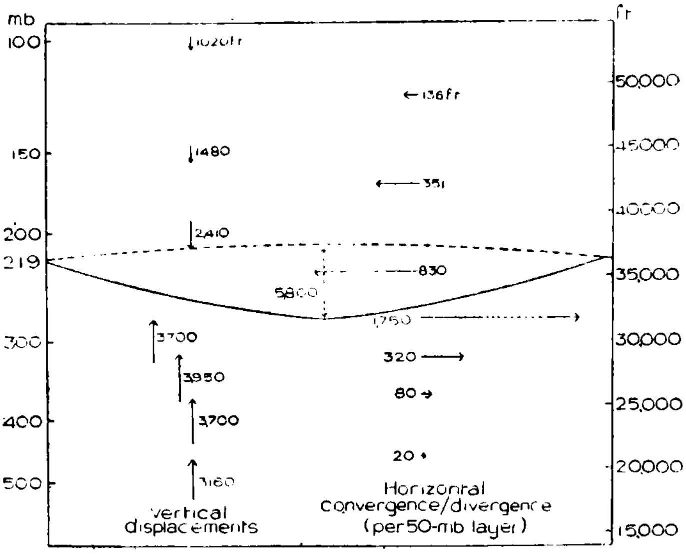


FIG. 4—CHANGE OF TROPOPAUSE FROM 210 MB. TO 270 MB. DURING OCTOBER

(ii) The horizontal divergence in the troposphere and the convergence in the stratosphere each tend to a maximum as the tropopause is approached. Even allowing that the components of convergence and divergence may be small in relation to wind speed, this probably means that appreciable shear of wind exists at the tropopause when tropopause level is changing.

(iii) In a downward change of the tropopause, the stratospheric air near the tropopause is rising in temperature by subsidence, but losing some heat by radiation. In the tropospheric air just below the effects are reversed. It is here probably therefore that the exchange of heat by radiation reaches its maximum.

Summary.—On the hypothesis—for the adoption of which reasons are given—that the air in the lower stratosphere of middle and high latitudes is air which has drifted, with subsidence, from the upper troposphere of the tropics, computations are made of the percentage losses by radiation of the increase of temperature which should have been gained if the subsidence had been under adiabatic conditions. It is found that the gross adiabatic gain of temperature would require to have been offset by radiation losses of 62 per cent. in winter or 36 per cent. in summer to produce the average distributions of temperature found in temperate latitudes in the stratosphere up to the 18-Km. level. The rate of loss by radiation requires to be about half a degree Fahrenheit a day in winter and a quarter of a degree in summer.

Radiation losses equalling or exceeding the adiabatic gains would be required to account for winter polar extremes of temperature. Losses exceeding around 70 per cent. would make inevitable the return of stratospheric air to the troposphere in the polar winter. Assuming corresponding time rates of radiation losses for the shorter-period time-scale of the depressions of middle latitudes, estimates are made of the air displacements associated with rapidly changing tropopause level.

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NIGHT COOLING UNDER CLOUDY SKIES

By W. D. SUMMERSBY, B.Sc.

Saunders[1], recently investigated night cooling under clear skies at Northolt, and on the basis of his results devised a practical routine for forecasting the night minimum temperature. The following is an attempt to arrive at a method for extending his work so that cloudy, or partly cloudy, nights also may be dealt with.

Method.—It was decided to select a number of cases at Northolt, ascertain the night minimum temperature as predicted in each case by Saunders's method on the assumption of no cloud, and then to determine the correction to be applied to this night minimum to bring it into agreement with the actual night minimum for each cloudy night. It was hoped by this means to discover a relation between the correction to the night minimum and the average cloud conditions during the night which could be used in practical forecasting.

In selecting suitable cases for this treatment, the following were excluded for obvious reasons:—

(i) Nights with no cloud or only cirrus.

(ii) Nights with any precipitation.

(iii) Nights during which any front, however weak, crossed the London area, or on which any advective change of dew point was noted. In this latter connexion, the normal tendency is for the dew point to rise slightly until the time of T_r (Saunders's notation for the screen temperature at the evening discontinuity in cooling rate) and then to fall slowly by some 3–4°F.[2] Nights on which this general tendency was shown were regarded as acceptable.

The cloud-amount parameter adopted was merely the arithmetic mean, A, of the number of oktas of cloud reported in the hourly observations, irrespective of cloud type or height. Here, following Saunders, fog with sky obscured was regarded as 8 oktas cloud, whilst in cases of fog with sky discernible, the cloud amount reported was used as with no fog. The minimum temperature as predicted for no cloud was subtracted from the observed minimum in each case yielding the correction (ΔT) which would be required. Simultaneous pairs of values of ΔT and A were plotted in the form of a dot diagram and the

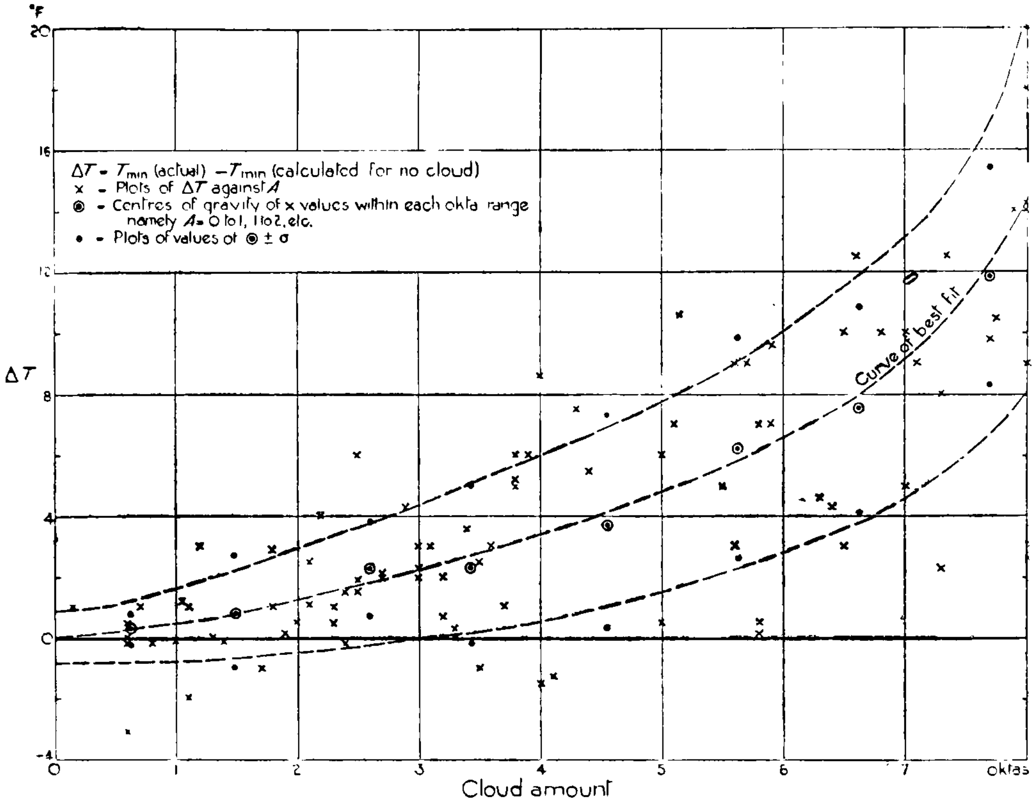

FIG. 1—RELATION BETWEEN AVERAGE CLOUD AMOUNT AND DECREASE IN COOLING

curve of best fit drawn; 90 cases were used for this diagram which is given at Fig. 1.

Discussion.—The curve of best fit was drawn simply through the centres of gravity of each group of points within the cloud-amount ranges: >7 oktas up to 8 oktas; >6 oktas up to 7 oktas, etc. As a check, the deviation, χ, of each point from the curve was evaluated, and an histogram drawn of the frequencies of deviations within given ranges, Fig. 2. The histogram shows central distribution, thus confirming the accuracy of the curve of best fit.

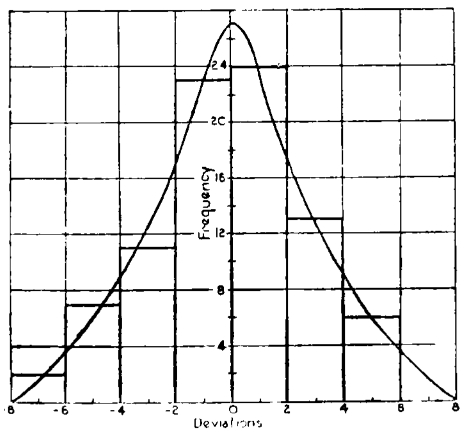

FIG. 2—FREQUENCY DISTRIBUTIONS OF DEVIATIONS AT ΔT FROM CURVE OF BEST FIT IN FIG. 1

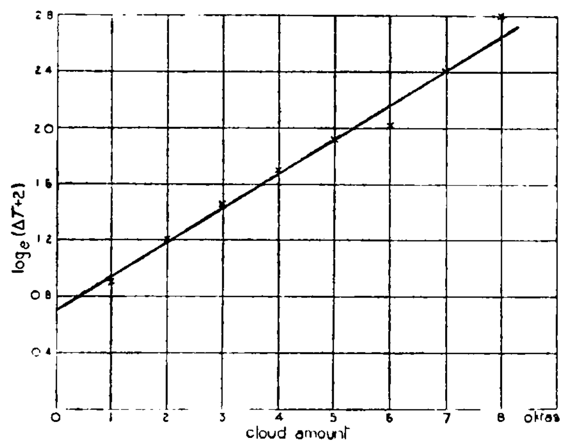

FIG. 3—GRAPH OF LOG$_e$ $(\Delta T + 2)$ PLOTTED AGAINST CLOUD AMOUNT

Having noted χ within each okta range of A, the root mean square deviation for each range was calculated, namely $\sigma = \sqrt{(\Sigma\chi^2/n)}$. The results are given in Table I.

TABLE I

Cloud amount range Greater than	Up to	σ	Cloud amount range Greater than	Up to	σ
oktas	oktas	°F.	oktas	oktas	°F.
7	8	3·59	3	4	2·60
6	7	3·23	2	3	1·52
5	6	3·59	1	2	1·84
4	5	3·46	0	1	0·56

The values of σ were used to draw the two curves shown above and below the curve of best fit in Fig. 1. As about 70 per cent. of the points have less deviation from the mode than the probable error, the curves enclose the area within which it is reasonable to expect about 70 per cent. of points to lie, and therefore give an indication of the accuracy which would be expected from using the curve of best fit in practice.

Relations of an exponential form are commonly encountered in dealing with cooling problems, and the shape of the curve obtained suggests a relation of the type:—

$$\Delta T = C_1\{\exp(C_2A) - 1\}$$

where C_1 and C_2 are constants

or $\log_e(\Delta T + C_1) = C_2A + \log_e C_1$.

If such a relation be assumed to apply, calculation from the data yields a value for C_1 of about 2°F. The result of plotting the values of $\log_e(\Delta T + 2)$ for the intercepts of the curve of best fit with the ordinates at $A = 1, 2, 3$, etc. against those values of A is shown in Fig. 3; the conspicuously linear nature of the curve will be noted. The slope of the line is found to be 0·246, so that to a close degree of approximation

$$\Delta T = 2(e^{A/4} - 1).$$

Table II, shows the agreement between the values of ΔT at $A = 1, 2, 3$, etc. taken from the curve of best fit and those calculated from the above formula.

TABLE II

A	ΔT curve	$\Delta T = 2(e^{A/4} - 1)$	A	ΔT curve	$\Delta T = 2(e^{A/4} - 1)$
0	0·0	0·00	5	4·8	5·00
1	0·5	0·58	6	6·5	6·96
2	1·3	1·29	7	9·1	9·53
3	2·2	2·25	8	14·3	12·78
4	3·4	3·44			

A further idea of the accuracy of the relation found was obtained by evaluating $2(e^{A/4} - 1)$ for each of the 90 cases, and finding the deviation, d, of each from the observed value of ΔT. The standard deviation for the whole okta range ($= \sqrt{(\Sigma d^2/90)}$) was found to be 2·74°F. This may be regarded as encouraging, especially as only cloud amount was taken into account.

It will be remembered that the standard deviations given in Table I vary from 0·56°F. for cloud amount 0 to 1 okta up to 3·59°F. for 7 to 8 oktas. This increase is clearly due to the greater effect of cloud type, height and thickness for large average cloud amounts than for small amounts. It was appreciated that these factors would be significant, but attempts to obtain separate curves corresponding to different cloud types and height ranges proved fruitless owing to the complications involved. There was a suggestion, however, that cloud thickness would be a more useful parameter to introduce as a refinement, and it is proposed to examine this question separately. In using these results in practice, a forecaster would of course depart from the curve of best fit for large cloud amounts, moving towards the upper curve in conditions of thick low cloud and to the lower curve with thin high cloud. In practice, therefore, greater accuracy might be obtainable with large cloud amounts than the figure 3·59°F. would appear to indicate.

Conclusion.—The results described provide the forecaster with a practicable method of forecasting the night minimum temperature on cloudy or partly cloudy nights, by applying a correction to the minimum as predicted by Saunders's method for clear skies[1]. Either the curve in Fig. 1 or tables based on the exponential relation may be used to find the correction. It is presupposed for this purpose that the average cloud amount for the night can be forecast successfully. The curve and relation obtained for ΔT are, strictly, applicable to Northolt alone but would probably be satisfactory at stations with similar subsoil and topography. At other stations a trial on similar lines would be desirable.

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METEOROLOGICAL OFFICE DISCUSSION

Fog investigations

The discussion on March 16, 1953, held at the Royal Society of Arts was opened by Dr. K. H. Stewart who described three groups of fog investigations undertaken by the Meteorological Office. He dealt first with the measurement of visibility at airfields; in the past there had been many complaints from pilots that meteorological reports of visibility did not give a true picture of the conditions they experienced on landing, and a good deal of work had been done to clear up the discrepancies. There were three obvious reasons for differences between pilots and ground observers:—

(i) Difference of place; the runway in use might be a mile or more from the place where routine meteorological observations were made.

(ii) Difference in object looked at; the meteorological quantity "visibility" was intended to be a measure of the properties of the atmosphere only, and was therefore defined and measured as the range at which ideal black objects disappeared against a sky background, whereas pilots, of course, were interested in the range at which the runway, or its lights, disappeared.

(iii) Difference in height; meteorological observations were normally made about 5 ft. above ground level, but a pilot might be looking down from a height of a few hundred feet.

The Runway Visual Range Scheme, in which observers close to the runway measured the range of the actual markers of lights used as guidance by pilots, had been introduced eighteen months ago and seemed to be very successful in overcoming the difficulties mentioned in (i) and (ii). Meanwhile experiments in the measurement of "slant visibility" to avoid the difference (iii) had been pursued. Dr. Stewart described the four methods of measuring slant visibility—by balloon-borne light, by pyrotechnic flares, by balloon-borne photo-electric fog-density indicator, and by inclined searchlight beams—that had been developed, and the attempts to test them by comparison with direct observations from aircraft. He said, however, that there was now some doubt as to whether slant-visibility measurements would really provide any extra useful information. Runway visual range provided a very good approximation to what pilots wanted to know, and it might well be that the inevitable errors due to the inherent patchiness of fog were more serious than those due to neglect of the vertical variation of fog density. The main aim of the work in the past winter had been to collect information about the time and space variation of fog density (by recording photo-electric visibility meters and by balloon-borne apparatus at Cardington) to help in deciding whether routine slant-visibility measurements were necessary at airfields.

The second set of investigations was that being made at Cardington. On occasions when radiation fog was expected, observations of temperature, humidity and wind speed up to 4,000 ft. were made, using apparatus hung from a large barrage balloon, in addition to normal surface observations. These observations should, in due course, lead to improved methods of fog forecasting, though it might well be several years before sufficient data are accumulated to give any significant improvement over methods such as that due to Saunders[1]. The Cardington observations could also provide a basis for more detailed experiments on particular physical processes. The first aim of these experiments was to find how and why the water content of the air near the ground decreased on a radiation night. The most obvious explanation was that it appeared as dew, and methods of measuring dewfall were being developed. More complicated measurements, such as those of radiation and convection, might be necessary later on. Problems involving the size and number of drops were also being considered, and measurements of drop-size distribution were planned.

Finally, Dr. Stewart showed slides illustrating the extent and intensity of fog in the London area in early December 1952. These had been prepared from reports collected from many sources in addition to ordinary synoptic stations. While the dirt and deadliness of the fog had been due to man-made pollution, he thought that the direct effect of solid polluting particles on visibility had been relatively small. Pollution might well have had important indirect effects on visibility, however, by providing many nuclei for condensation. The fog had therefore probably contained a larger number of very small droplets than most fogs, and the chances of partially clearing it by means of a brine spray would have been greater than usual.

Mr. Rustom, of Pan American Airways, said he thought that runway visual range measurements would be improved if brighter lights were used in the estimation of range; on some occasions of patchy fog the runway lights could be seen through the fog while the goose-neck flares used in range estimation could not. He felt that slant-visibility information would save many abortive approaches.

Mr. Wallington also thought there was a need for pilots to know the slant visibility before they started to approach. Visibility on an approach was often only half that on the ground, particularly in smoky conditions. In forecasting fog, he thought it was necessary to allow some time for the fog particles to grow sufficiently to affect visibility, after the theoretical fog point was reached.

Mr. Harley quoted a recent case at London Airport where the visual range on the approach was far less than that on the ground because of the dazzling effect of the sun on a haze layer.

Mr. Mercer, of the Ministry of Supply Blind-Landing Experimental Unit, said that experience in fog trials had shown that pilots found advance information as to when they should pick up lights very helpful. This information could be given quite reliably from simple measurements with a balloon-borne light. In many cases conditions at 200 ft. had been very significantly worse than those at ground level. He did not think that the patchiness of fog was a good reason for limiting the accuracy of measurements; it would be better to measure the patchiness. He ended by expressing great appreciation of the co-operation of the staff of the Meteorological Office at Wattisham in providing forecasts for his trials.

Mr. Russell said that slant-visibility measurements would be useful in the development of FIDO.

Mr. Harley said that if, as had been suggested, it was safe for a pilot to descend to a critical height of about 200 ft. from which he could estimate the slant visibility, it seemed unnecessary to measure this quantity from the ground.

Mr. Stallibrass said that some estimate of slant visibility was essential for civil aviation purposes. He agreed that it might turn out that it was better to deduce the slant visibility indirectly from surface observations than to measure it directly.

Mr. Durward emphasized that slant visibility, like surface visibility, was a variable quantity.

Mr. Saunders suggested that it would be better to collect slant-visibility statistics near London than at Cardington; he thought that smoke had important effects on slant visibility. Referring to the question of loss of water on a radiation night, he thought that it was important to establish the normal mode of variation of dew point at various heights in the absence of advection.

Mr. Poulter said that he had investigated the fall of temperature close to the ground. The discontinuity reported by Saunders seemed to occur at a fixed time after sunset.

Mr. Jacobs (partly communicated) reported on some tests of Saunders's method of predicting fog point. At Defford, of the 47 fogs in the last 6 months only 11 could be classified as true water fogs. The calculated and actual fog points were the same, within 1°F., on 8 of these 11 occasions. In two of the remaining three cases the temperature fell 4–5°F. below the estimated fog point before a fog formed, but there was difficulty in finding representative radio-sonde ascents (a difficulty which occurs on other occasions at Defford) particularly as rain occurred before the sky cleared. On the last occasion the fog point was reached only for a short time. At Filton, fog formed on 24 of the 59 radiation nights during the last year and water fog always occurred. In 17 of these 24 cases the calculated fog point was within 2°F of the actual fog point (12 within 1°F.); in 4 of the remaining 7 cases the estimated fog point was below freezing point and air and ground frost, which did occur, complicated the issue. Of the 35 radiation nights when no fog formed, the calculated fog point was not reached on 21, and of the remaining 14 no less than 10 had a calculated fog point of below or near freezing point, frost actually occurring on all these occasions. Further tests are being made at stations in the Gloucester group of meteorological offices.

Cmdr Frankcom asked if it had been foreseen that the early December fog would be so persistent in the lower reaches of the Thames; he felt that forecasts for this area were important. In some cases, e.g. in the Tees–Tyne area, he thought that smoke alone could have a big effect on visibility. He suggested that counts of nuclei in the very pure air at an ocean weather station might be useful.

Mr. Illsley pointed out that visibility indoors during the London fog of early December, for example in Covent Garden Opera House, had been very low at times, and suggested this was presumably due to pollution alone, without water drops.

Dr. Scrase said he thought that measurements by Wright[2] at Kew had shown that the removal of smoke would produce a great improvement in visibility.

Mr. Clark said that the idea of using a salt spray to clear fog was well known, but had various disadvantages. Other methods, using electrical deposition or dispersal by heat were also available; during the war, the FIDO method appeared to be the cheapest.

Dr. Stewart, in reply, agreed with Mr. Rustom that brighter lights would be an advantage in the measurement of runway visual range. He believed that smoke pollution alone could reduce visibility to 200 or even 100 yd., but that pollution in the amounts actually measured during the December fog could not reduce it to lower values. The indirect effects of pollution, in providing nuclei, might be important and might explain the indoor observations if the relative humidity had been much above 70 per cent.

REFERENCES

1. SAUNDERS, W. E.; Method of forecasting the temperature of fog formation. *Met. Mag., London,* **79**, 1950, p. 213.
2. WRIGHT, H. L.; Atmospheric opacity: a study of visibility observations in the British Isles. *Quart. J. R. met. Soc., London,* **65**, 1939, p. 411.

ROYAL METEOROLOGICAL SOCIETY

At the meeting of the Society held on February 18, 1953, the President Sir Charles Normand in the chair, the following papers were read:—

*Best, A. C.—Condensation nuclei and the development of radiation fog**

Mr. Best said his paper was mainly an attempt to see if the size distribution of condensation nuclei, supposed to be sea salt, had any effect on the formation of fog as relative humidity increased from 80 to 100 per cent. It was assumed that as the humidity increased the size of the drops increased in equilibrium with the humidity of the surrounding air. This assumption was justified, except for large nuclei in nearly saturated air, by showing that the time taken for the drop size to respond to a sudden increase of 5 per cent. in relative humidity was less than 1 sec. Arbitrary size distributions of nuclei were assumed and the following quantities computed as relative humidity increased:

* *Quart. J.R. met. Soc., London,* **79**, 1953, p. 112.

(*a*) total water content
(*b*) visibility (from Koschmieder's and Houghton's formulae)
(*c*) drop-size distribution at saturation. It was found that the total water content and the atmospheric opacity in a developing fog were proportional to the salt content, but that the shape of the nucleus-distribution curve had little effect on them. The calculated visibility and drop-size distribution in air just become saturated agreed well with observation; as relative humidity increases to slightly over 100 per cent. the drops can grow indefinitely in size.

In the discussion some criticism was expressed by Mr. Ludlam and Mr. Gold of the validity of the assumption of equilibrium between drop size and humidity in nearly saturated air. Dr. Lessing discussed the effect of pollution of urban fogs by sulphur compounds which he believed made the fog more stable. Mr. Mason said little was known of the actual distribution of nuclei in the air and that observations at Los Angeles showed that relative humidity could vary rapidly in fog by several per cent. Mr. Sawyer said he had noticed that if the air was clear then, as relative humidity increased, any fog that formed did so suddenly, whereas in polluted air there was a gradual decrease in visibility.

*Saunders, W. E.—Some further aspects of night cooling under clear skies**

Mr. Saunders described a method of forecasting the temperature during a radiation night for purposes of fog prediction. He had noticed some years ago the existence of a sudden reduction after dusk in the rate of fall of temperature over grass in England. A similar effect was, he pointed out, noticeable in readings of temperature near the ground made in Texas, which showed that the discontinuity was very pronounced near the ground and was less pronounced in the screen. It had been suggested to him that the discontinuity might mark the onset of dew which provided a source of heat. A regression equation method of calculating the temperature of the discontinuity from the day maximum temperature and dew point was described, and curves were set out from which the rate of fall of temperature after the discontinuity could be determined. The method gives a prediction of the minimum temperature as well as the rate of cooling. For full details reference must be made to the paper*.

The discussion turned mainly on the reason for the discontinuity in the rate of fall. Dr. Penman believed it to be real and associated with the cessation at nightfall of transpiration from plants. Dr. Robinson could find no discontinuities in Kew thermograms but a very marked one in the record of net inward radiation flux at the ground. Dr. Sutcliffe said the paper was a very serious practical contribution to fog forecasting whatever the cause of the discontinuity. Mr. Gold said dew sometimes formed even before sunset. Mr. Pilsbury commented that the time of discontinuity changes rapidly after the first radiation night in October which suggested lowered soil temperature was of basic importance. Mr. Saunders in reply agreed with Mr. Gold that some dew might form before sunset, but he had observed copious dew formation at the time of the discontinuity, and that the discontinuities occurred in Texas at the same times relative to sunset as they did in England. In reply to a query by Mr. Bonacina he thought the reason for the increase in difference between screen and grass minima with decreasing temperature was a decrease in the amount of atmospheric moisture.

* *Quart. J.R. met. Soc., London,* **78**, 1952, p. 603.

LETTER TO THE EDITOR

Rainbow phenomenon

The rare phenomenon mentioned in Mr. Pilsbury's letter published in the *Meteorological Magazine* for February 1953 was observed at Benbecula on December 16, 1952. The arcs observed were travelling slowly through the straits separating Benbecula from North Uist.

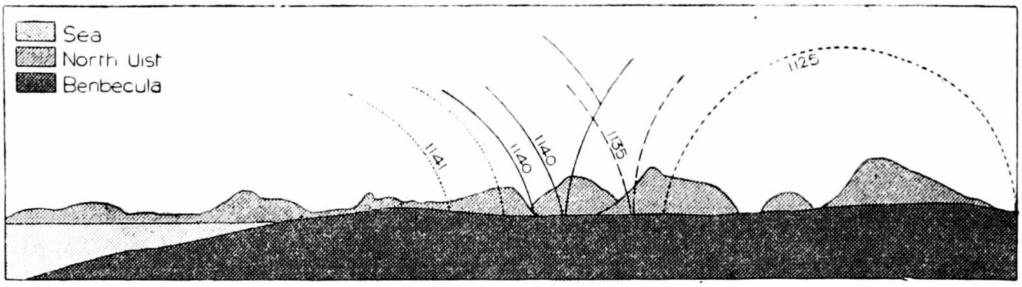

FIG. 1—LOOKING NORTH FROM CONTROL TOWER, BENBECULA AIRPORT

The weather had been cloudy with continuous moderate rain from 8 oktas of nimbostratus at 1,500 ft. with patches of fractostratus at 1,000 ft., the wind at the time being 180° 15 kt. At 1055 G.M.T. a break in the cloud became visible in the south-west, and at 1115 the wind veered to 280° 12–15 kt. indicating the passage of a trough. After this, the wind dropped slightly and the rain became intermittent, and during this period a faint but complete rainbow was observed at 1125 between Benbecula and North Uist. At 1135 an arc formed in the opposite direction to the original bow which had faded and now had only an arc composed of the western half of the bow, red being uppermost in both arcs. Subsequent developments are shown in Fig. 1, red being uppermost in all the arcs shown. In this diagram the lengths of the arcs and the angles between the bows are only approximations.

L. E. WATSON

Benbecula Airport, Outer Hebrides, March 2, 1953

NOTES AND NEWS

Hurricane force winds at ocean weather station "India"

During Voyage 42 of O.W.S. *Weather Watcher*, while on duty at station INDIA (59°N., 19°W.) on December 22, 1952, the vessel experienced winds of hurricane force. A heavy sea struck the ship and completely smashed her two starboard boats. The ship returned to Londonderry forthwith, and, having landed her two smashed boats, borrowed two similar boats from *Weather Observer* who was at that time homeward bound from station JULIETT. It is unusual for one of these ships to take any heavy water on board. In fact, this is the first occasion that any of the weather ships has sustained sea damage worth mentioning in the 4½ years they have been in operation.

The following extracts from the Master's report upon this occurrence show the way in which this exceptionally heavy sea built up:—

"The weather from midnight 21st to midnight 22nd December, 1952, was as follows:

Midnight	21st	Wind ESE., having just freshened to force 8–9, sea 12 ft.
0100	22nd	Wind ESE., force 11–12, sea 15 ft.
0200		Wind ESE., force 10–11, sea 18 ft.
0300		Wind SE., force 10, sea 20 ft.
0400		Wind S'ly, force 10, sea 18 ft.
0800		Wind S'W., force 9, sea 20 ft.
1200		Wind SSW., force 11–12, sea 25–30 ft.
1300		Wind SSW., force 12, sea 28–30 ft. Very confused.
1600		Wind SW., force 12, sea 28 ft. Becoming less confused.
2000		Wind SW., force 12, sea 30–35 ft., with occasional 40-ft. seas but now true in direction.
Midnight	22nd	Wind SW., force 10–11, sea 28 ft.

Wind continued SW'ly, moderating steadily to force 8 by 0800 23/12/52.

"At midnight December 21, 1952, the ship had been stopped for a number of hours, wind at the time being ESE. force 8–9 and increasing, seas 12 ft. By 0100 the wind had increased to force 11–12, and the seas were 15 ft. Ship was now rolling very heavily so we commenced steaming into wind and sea at 60 revolutions, minimum speed to give steerage way. As the seas became higher speed had to be steadily increased and by 0800 we were doing 70 revolutions, and at 1130 we further increased speed to 75 revolutions. Our courses were, at 0530 185 degrees, at 0700 190 degrees, and at 1130 195 degrees.

"Ship was riding the seas well and we had not shipped any really heavy water, nor did I consider we would under the present conditions.

"At 1230 when we were comfortably having dinner the ship suddenly gave a violent lurch to port and there was a loud report of a sea striking the ship. I immediately went up on deck to find that we had shipped what must have been an exceptionally heavy sea and that our starboard lifeboat and dinghy had been smashed beyond repair, as well as other damage."

C. E. N. FRANKCOM

REVIEW

Micrometeorology. By O. G. Sutton. 9 in. × 6 in., pp. xii + 333, *Illus.*, McGraw-Hill Publishing Co. Ltd, London. Price 61*s.* net.

To many meteorologists the subject of turbulence in the lowest layers of the atmosphere has some of the characteristics of a "closed shop". The complaint has been heard that the subject has its own jargon, and has sometimes been accompanied by the implication that the devotees are only related to the family of meteorologists and are not part of it. To some extent this attitude probably arises from the lack of a comprehensive and connected account of the subject (in English). This lack has now been remedied.

Professor Sutton's declared intention has been "to meet the needs of meteorologists, and of workers in other fields, who require detailed information about physical processes in the regions of the atmosphere where life is most abundant". To meet these needs the author has written a book of which roughly the first half provides the basic physics and mathematics and the second half gives a detailed discussion of the extent to which experiment supports, or refutes, theory. Throughout the early pages the development of the basic mathematics and physics is carried out with a view to application to the lower atmosphere. The result is a connected and self-contained textbook, suitable alike for the student about to start work in this field and for the research worker who requires "an integrated account of the major developments in the field". Inevitably the subject is mathematical, but in only a few places will the reader of modest mathematical ability find any real difficulty, and then only if he is not prepared to accept the stated solution to a differential equation.

After a preliminary short first chapter dealing with the usual relations governing temperature and humidity in a static atmosphere the second chapter treats laminar flow in the atmosphere. The equations of motion of an inviscid fluid are developed and applied to irrotational air flow over an undulating surface and over a cliff and to winds on a rotating earth. The Navier-Stokes equations for motion of an incompressible viscous fluid are obtained and applied to flow over a plane surface. Finally, the idea of a boundary layer in laminar flow and the appropriate equations are developed and applied to flow over a plane surface.

Chapter 3 deals with turbulent flow. The Reynolds' stresses are introduced, and the idea of exchange coefficients developed by analogy with molecular processes and applied to the problem of the approach to the geostrophic wind. The author then introduces the idea of a turbulent boundary layer over smooth and rough surfaces, and follows with a consideration of velocity profiles in this layer in relation to the mixing length hypothesis. The chapter finishes with a discussion of statistical theories of turbulence and a brief section devoted to the similarity theory.

Chapter 4 discusses heat transfer and problems of diffusion. The mathematics of the diffusion of heat in a fluid and the idea of a thermal boundary layer are followed by a section which discusses free convection as a result of buoyancy. A long section then treats the solution of the Fickian diffusion equation, with a constant transfer coefficient, for various different boundary conditions corresponding to different meteorological applications. The mathematics of turbulent heat transfer, both by forced and free convection and in relation to the mixing length concept, are discussed, and the chapter closes with a discussion of the Richardson number as a criterion of the effect of density gradient on the growth of turbulence.

Chapter 5 deals with radiation. Following a brief description of the fundamental relations there are factual descriptions of the passage of short-wave radiation to the earth and of the transfer of heat by long-wave radiation. The last part of the chapter discusses the factors governing the night-minimum temperature and various formulae which have been suggested for prediction of this temperature.

Chapter 6 is entitled "Temperature field of the lowest layers of the atmosphere". It starts with a description of surface conditions, including temperature variations in the ground, the effect of soil characteristics and of soil cover and the variation of surface temperature. Then the published measurements of temperature profiles in the lowest layers are discussed and analysed. There follows a theoretical treatment of heat transfer in the lowest layers of the atmosphere leading to the variation of temperature and to the propagation of the diurnal temperature wave in the vertical. This is done for both constant diffusivity and for a diffusivity which varies as a power of the height, and is accompanied by a comparison with observations of the variation with height of the phase and amplitude of the diurnal wave. The chapter finishes with a short discussion of the effects of buoyancy and of radiative transfer of heat.

Chapter 7 discusses problems of wind structure near the surface. The available data for the first few metres are described in terms of the logarithmic profile for neutral stability and on the basis of Deacon's formula for other stabilities. The author then discusses the approach to the geostrophic wind, following Köhler, with an eddy coefficient represented by a power of the height. A detailed discussion of wind near the surface deals with eddy velocities, gustiness, the drag of the earth's surface and the Reynolds' stresses and relates the measurements to wind-velocity profiles. After a brief section on the effect of stabilizing density gradients the chapter closes with a comparison of the theory of winds on slopes with observations.

The final chapter discusses diffusion and evaporation. Following a description of the characteristics of clouds from artificial sources the theoretical solutions of the diffusion equation for various kinds of source and for constant and variable eddy coefficients are given. The solutions obtainable using the statistical theory of turbulence are examined. This is followed by a brief discussion of diffusion from an elevated source and a discussion of the free convective jet in the atmosphere. Evaporation from a saturated surface is treated primarily as a problem in diffusion for both limited and unlimited areas and a comparison made with experimental results. Finally there are brief sections on a comparison of the exchange coefficients for heat, momentum and matter and on the similarity theory of turbulence.

Each chapter is accompanied by a bibliography.

Inevitably the present book invites comparison, or contrast, with the monograph on "Atmospheric turbulence" by the same author. Although there is much similarity in the subject matter the latest book is not simply an expanded version of the monograph. "Micrometeorology" is a course of reading, planned as a continuous treatment and designed to carry the student from the hydrodynamics of a frictionless fluid to our present knowledge of the structure of the lower atmosphere. This has involved some repetition but this is not a disadvantage. Throughout the book theory and experiment are related, but an essential feature is the change in emphasis in about the middle from the basic work of the first four chapters to observed facts and their relation to theory in the last three chapters.

There is no doubt that the author's reputation will be greatly enhanced by his latest contribution to meteorology. It very adequately fills an obvious gap in the literature, and is one of the most important meteorological textbooks produced in English during the past twenty years. The production is excellent and the reviewer noticed very few misprints.

A. C. BEST

OBITUARY

Clarence Vivian Davies Bolton.—It is with great regret that we record the death, on April 14, 1953, at the age of 46, of Mr. C. V. D. Bolton, Experimental Officer, Llandow.

Mr. Bolton joined the Office in March 1937, and served as assistant and, later, forecaster mainly at Royal Air Force stations in the United Kingdom. From 1950 until his last illness he was in charge of the meteorological office at St. Athan, later transferred to Llandow. He was a loyal, cheerful and efficient officer whose first interest was his work. Mr. Bolton left a widow, a son and daughter, to whom we offer our sincerest sympathy in their great loss.

HONOURS

The following awards were announced in the Coronation Honours List, 1953:

C.M.G.

Mr. J. Durward, Deputy Director (Services), Meteorological Office

O.B.E.

Mr. A. C. Best, Principal Scientific Officer, Meteorological Office

OFFICIAL ANNOUNCEMENT

Appointment of Deputy Directors of the Meteorological Office.—Mr. S. P. Peters has been appointed Deputy Director for Forecasting in succession to Mr. E. G. Bilham, and Dr. R. C. Sutcliffe, O.B.E., Deputy Director for Research in succession to Dr. A. H. R. Goldie, C.B.E.

The appointments took effect on June 1, 1953.

METEOROLOGICAL OFFICE NEWS

Ocean weather ships.—The following is an extract from the report of the Master, *Weather Observer*, Voyage 47: "This has been the best voyage I have experienced, the weather for almost all the period on station (April 5–25, 1953) has been perfect with the crew sunbathing and swimming almost daily. During the last week on station (52°30′N. 20°00′W.) a number of swallows and martins flew around the ship in an exhausted state; some were caught and the Second Officer tried valiantly to feed the birds with a fountain-pen filler but unfortunately they died."

Sport and Athletics.—*Bishop Shield.*—On May 29, at Chiswick, for the fifth year in succession, the Meteorological Office Harrow Ladies' netball team won the Air Ministry Bishop Shield competition.

Walking.—Mr. G. M. Band is evidently maintaining his form: he won the Iraq Command Walking Championship held earlier in the year at Habbaniya, covering the seven-mile course in 59 minutes.

WEATHER OF MAY 1953

Mean pressure was above normal in the region extending south-eastward from Greenland to Iceland and most of Europe except the south; it was below normal over the North Atlantic south of latitude 60°N. and over most of North America. The greatest excess of pressure, 6 mb., occurred in the region of north-east Greenland, where mean pressure was 1019 mb. while the greatest deficit of pressure, also 6 mb., was at approximately 50°N. 30°W. where mean pressure fell to 1008 mb. At the Azores the mean pressure was 4 mb. below normal.

Mean temperature was above normal in Europe, generally between 2° and 5°F., and below normal over most of the United States except the east. The mean temperature varied from 45° to 50°F. in Scandinavia, 55° to 65°F. in west and central Europe and 65° to 70°F. in the Mediterranean region.

In the British Isles the weather was rather warm generally and sunny in most districts; it was dry until the 13th. A very warm spell occurred in England and Wales during the Whitsun holiday on the 24th and 25th. Severe thunderstorms in northern and central districts of Great Britain on the 25th caused floods and considerable damage in places, with some loss of life by lightning.

An occlusion over the southern North Sea gave rain in the south-east on the 1st. Meanwhile an anticyclone built up off the south-west coasts and moved north-east over the country, becoming more intense. This system maintained fine weather almost everywhere until the 13th. There were fairly widespread early morning ground frosts up to the morning of the 4th and day temperatures reached or somewhat exceeded 70°F. in places from the 3rd to the 6th. On the 5th and 6th the anticyclone moved north to the Norwegian Sea but pressure continued high in a wedge across the British Isles, and fine weather persisted although it was generally cooler. Ground frost was again rather widespread in the early morning from the 9th to the 13th. A spell of unsettled weather ensued. On the 13th and 14th a deep Atlantic depression approached west Ireland and an associated trough of low pressure moved north-east over the British Isles, giving rain in southern districts on the 13th and throughout the country on the 14th. Subsequently the main depression moved north-east, and further secondary disturbances crossed the country giving rain or showers and local thunderstorms. On the 18th and 19th a small depression moved northward from south-west France to east Scotland giving general rain and thunderstorms at some places on or near the south-east coast of England. Thereafter pressure was relatively high over France, while Atlantic depressions moved north-north-east along the western seaboard and associated troughs gave rain in most areas (2·20 in. at Maesteg, Glamorganshire on the 22nd) but it was only slight in the south-east. By the 23rd a warm moist air stream of tropical origin covered most of the country with some fog on the south and west coasts. An anticyclone built up over France and moved north-east; south-easterly winds set in over most of the British Isles and very warm weather prevailed over England and Wales, with temperatures up to 80°F. at some places in the south-east on the 24th and over much of England on the 25th, when London Airport registered a maximum of 89°F. On the 24th a cold front, associated with a small depression which developed off north-west Scotland, approached west Ireland and subsequently moved east giving widespread, severe thunderstorms in many northern and central districts; the storms were accompanied by heavy rain and floods (2·52 in. at Fort William during the 24 hours ended at 0900 on the 25th, 2·12 in. at Barnard Castle, Durham, on the 25th) and several people were killed or injured by lightning. There were thunderstorms in east and south-east England on the 27th. The rest of the month was mainly dominated by a large anticyclone centred to the south-west of Ireland and rather cool, changeable weather prevailed, though rainfall amounts were mainly small.

The general character of the weather is shown by the following provisional figures.

	AIR TEMPERATURE			RAINFALL		SUNSHINE
	Highest	Lowest	Difference from average daily mean	Per-centage of average	No. of days difference from average	Per-centage of average
	°F.	°F.	°F.	%		%
England and Wales ...	89	26	+2·6	109	−3	113
Scotland	79	24	+2·6	108	−2	98
Northern Ireland ...	75	31	+3·3	90	−3	107

RAINFALL OF MAY 1953

Great Britain and Northern Ireland

County	Station	In.	Per cent. of Av.	County	Station	In.	Per cent. of Av.
London	Camden Square ...	1·41	80	*Glam.*	Cardiff, Penylan ...	3·47	142
Kent	Dover	1·73	104	*Pemb.*	Tenby, The Priory ...	2·84	123
,,	Edenbridge, Falconhurst	1·52	82	*Radnor*	Tyrmynydd	3·19	93
Sussex	Compton, Compton Ho.	3·24	146	*Mont.*	Lake Vyrnwy	3·23	100
,,	Worthing, Beach Ho. Pk.	2·09	127	*Mer.*	Blaenau Festiniog ...	5·55	98
Hants.	Ventnor Cemetery ...	2·51	144	,,	Aberdovey	1·78	71
,,	Southampton, East Pk.	2·65	133	*Carn.*	Llandudno	1·77	99
,,	South Farnborough ...	1·83	105	*Angl.*	Llanerchymedd ...	3·21	137
Herts.	Royston, Therfield Rec.	1·48	76	*I. Man*	Douglas, Borough Cem.	2·63	105
Bucks.	Slough, Upton... ...	1·29	77	*Wigtown*	Newton Stewart ...	2·52	95
Oxford	Oxford, Radcliffe ...	1·68	90	*Dumf.*	Dumfries, Crichton R.I.	2·37	86
N'hants.	Wellingboro' Swanspool	1·42	73	,,	Eskdalemuir Obsy. ...	2·90	88
Essex	Shoeburyness	1·49	115	*Roxb.*	Crailing	1·58	79
,,	Dovercourt	1·44	104	*Peebles*	Stobo Castle	2·41	106
Suffolk	Lowestoft Sec. School...	1·14	71	*Berwick*	Marchmont House ...	2·39	97
,,	Bury St. Ed., Westley H.	1·62	89	*E. Loth.*	North Berwick Res. ...	2·35	118
Norfolk	Sandringham Ho. Gdns.	1·43	78	*Midl'n.*	Edinburgh, Blackf'd. H.	1·79	87
Wilts.	Aldbourne	2·29	116	*Lanark*	Hamilton W. W., T'nhill	1·70	71
Dorset	Creech Grange... ...	3·09	151	*Ayr*	Colmonell, Knockdolian	1·93	75
,,	Beaminster, East St. ...	3·73	181	,,	Glen Afton, Ayr San. ...	..	..
Devon	Teignmouth, Den Gdns.	3·48	190	*Renfrew*	Greenock, Prospect Hill	3·04	93
,,	Cullompton	2·27	105	*Bute*	Rothesay, Ardencraig...	2·48	82
,,	Ilfracombe	3·07	149	*Argyll*	Morven (Drimnin) ...	3·64	113
,,	Okehampton	4·05	128	,,	Poltalloch	5·35	185
Cornwall	Bude, School House ...	2·45	133	,,	Inveraray Castle ...	3·89	99
,,	Penzance, Morrab Gdns.	2·87	130	,,	Islay, Eallabus... ...	3·57	135
,,	St. Austell	4·12	170	,,	Tiree	2·30	92
,,	Scilly, Tresco Abbey ...	2·83	167	*Kinross*	Loch Leven Sluice ...	2·47	101
Glos.	Cirencester	2·61	127	*Fife*	Leuchars Airfield ...	1·66	85
Salop	Church Stretton ...	2·64	104	*Perth*	Loch Dhu	5·72	127
,,	Shrewsbury, Monksmore	2·03	104	,,	Crieff, Strathearn Hyd.	2·36	95
Worcs.	Malvern, Free Library...	2·79	129	,,	Pitlochry, Fincastle ...	3·12	147
Warwick	Birmingham, Edgbaston	2·96	138	*Angus*	Montrose, Sunnyside ...	1·94	95
Leics.	Thornton Reservoir ...	2·05	102	*Aberd.*	Braemar	2·62	110
Lincs.	Boston, Skirbeck ...	2·62	149	,,	Dyce, Craibstone ...	2·62	103
,,	Skegness, Marine Gdns.	2·23	131	,,	New Deer School House	1·79	82
Notts.	Mansfield, Carr Bank...	..	..	*Moray*	Gordon Castle	1·44	68
Derby	Buxton, Terrace Slopes	2·43	78	*Nairn*	Nairn, Achareidh ...	3·07	172
Ches.	Bidston Observatory ...	2·02	106	*Inverness*	Loch Ness, Garthbeg ...	3·88	156
,,	Manchester, Ringway...	1·83	86	,,	Glenquoich	6·35	116
Lancs.	Stonyhurst College ...	2·46	86	,,	Fort William, Teviot ...	5·81	147
,,	Squires Gate	2·09	100	,,	Skye, Duntuilm ...	2·61	92
Yorks.	Wakefield, Clarence Pk.	1·80	91	,,	Skye, Broadford ...	4·18	99
,,	Hull, Pearson Park ...	2·30	119	*R. & C.*	Tain (Mayfield) ...	2·87	139
,,	Felixkirk, Mt. St. John...	1·84	98	,,	Inverbroom, Glackour...	3·18	106
,,	York Museum	1·74	87	,,	Achnashellach	4·84	114
,,	Scarborough	1·81	95	*Suth.*	Lochinver, Bank Ho. ...	2·50	98
,,	Middlesbrough... ...	1·62	84	*Caith.*	Wick Airfield	2·56	124
,,	Baldersdale, Hury Res.	2·61	105	*Shetland*	Lerwick Observatory ...	1·91	91
Norl'd.	Newcastle, Leazes Pk....	2·00	101	*Ferm.*	Crom Castle	2·99	107
,,	Bellingham, High Green	2·78	116	*Armagh*	Armagh Observatory ...	1·40	59
,,	Lilburn Tower Gdns. ...	2·32	100	*Down*	Seaforde	2·68	102
Cumb.	Geltsdale	2·60	101	*Antrim*	Aldergrove Airfield ...	1·63	72
,,	Keswick, High Hill ...	2·49	78	,,	Ballymena, Harryville...	1·73	60
,,	Ravenglass, The Grove	2·54	91	*L'derry*	Garvagh, Moneydig ...	2·38	93
Mon.	A'gavenny, Plâs Derwen	3·08	104	,,	Londonderry, Creggan	2·54	97
Glam.	Ystalyfera, Wern House	4·46	128	*Tyrone*	Omagh, Edenfel ...	3·44	133

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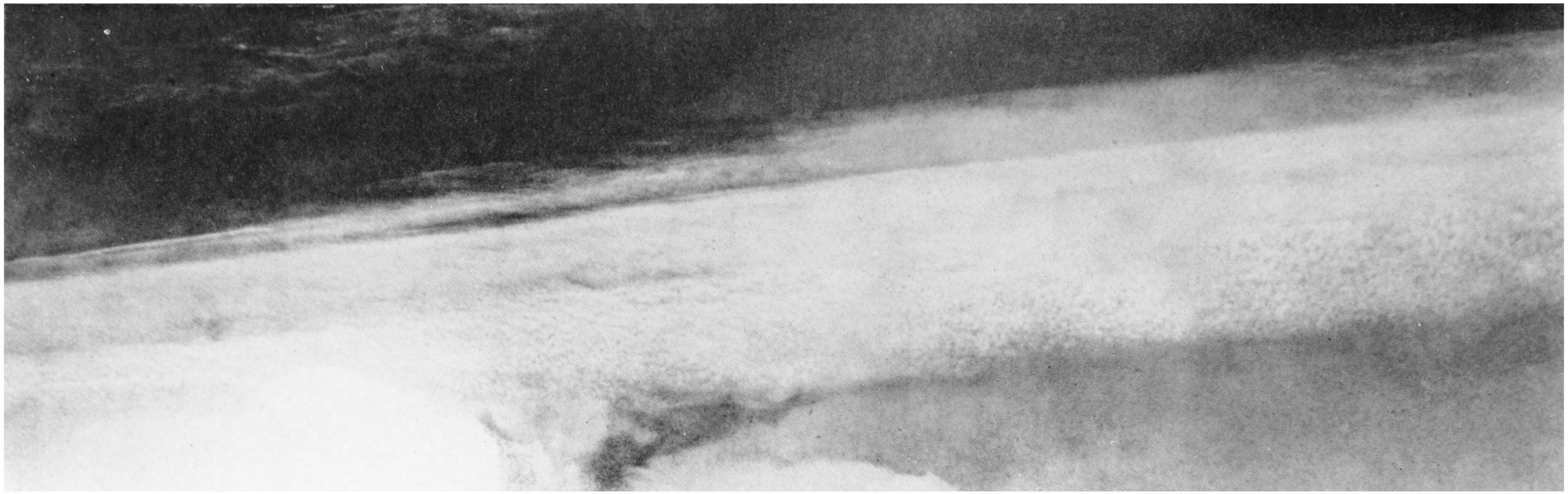

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CIRRUS CLOUDS, HARLINGTON, MIDDLESEX, MAY 14, 1953, 1400 G.M.T.

The line of cirrus photographed was one of a number which formed and dissipated throughout the day in the northerly air stream which covered the British Isles. The clouds formed at approximately 20,000 ft., and extended north–south for about 43 miles. These clouds are of the type considered by Mr. F. H. Ludlam to be of orographic origin (see p. 52).

Photo by the Royal Aircraft Establishment, South Farnborough

CREPUSCULAR RAYS