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APPLICATION OF AN INSTABILITY INDEX TO REGIONAL FORECASTING

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Introduction—Showalter[1] and Galway[2], working in the United States of America, introduced the "stability index" and the "lifted index," both relatively simple parameters to be used as aids in predicting local storms. The Showalter stability index is a static measure of latent instability, computed by lifting a parcel or bubble of air adiabatically from 850 millibars to 500 millibars. The theoretical temperature of the lifted parcel is then subtracted algebraically from the environment temperature at 500 millibars. Positive numbers indicate stability and negative numbers, instability. The lifted index is computed by similar but less objective methods, making use of the forecast maximum temperature. Using a similar technique, the author has attempted to produce a simple parameter to be called the instability index, for use in the production of regional forecasts when conditions are favourable for the development of air-mass type thunderstorms. The intention has been to provide an index which can be rapidly computed from the 2300 GMT temperature soundings and plotted on a small-scale chart. Isopleths may then be inserted to delineate areas of maximum latent instability and the chart can be used in compiling forecasts which have to be prepared early in the day, often before 0600 hours. Although some values of the index are as liable to precede showers as not, it is felt that the use of the method set out below will assist forecasters to delineate the most probable areas of thunderstorm occurrence and will also prevent failure to forecast thunderstorms on some occasions, as illustrated by the examples.

Computation of the instability index.—A network of thirteen radiosonde stations was used, nine in the British Isles and four adjacent continental stations, namely De Bilt, Uccle, Trappes and Brest. It was impracticable to make a rapid assessment of the theoretical maximum temperatures for all thirteen locations, extending as they do over a relatively large area, with varying degrees of exposure to maritime influences, and it was therefore decided to use an entirely objective method, in order to produce a working chart as quickly as possible. With this aim in view, the author used the 900-millibar wet-bulb potential temperature, since this measurement, whilst being representative of the air at low levels, would not be affected to any degree at night by outgoing terrestrial radiation. At pressures higher than 900 millibars, temperature and humidity fluctuations arising from nocturnal radiation, stratification and condensation on nuclei or the earth's surface, will be reflected in the lowest readings of the midnight soundings. Normal convention was followed in taking the 500-millibar dry-bulb temperature as the second reading to be used in the calculation, since this reading is indicative of the thermal structure in the middle troposphere and usually reflects the warm or cold tongues of the thickness chart. The 500-millibar dry-bulb temperature was subtracted algebraically from the 900-millibar wet-bulb potential temperature (both measured in degrees Celsius), the result being the instability index. Expressed as a simple formula,

$$\Delta T = \theta_{w900} - T_{500}$$

where, ΔT is the instability index

θ_{w900} is the 900-millibar wet-bulb potential temperature

T_{500} is the 500-millibar dry-bulb temperature.

The instability index for each upper air sounding can be calculated in a few seconds from the tephigram. Since dew-point is reported and not wet-bulb temperature, the intersection of the dry-adiabatic and mixing-ratio lines, appropriate to the 900-millibar dry-bulb and dew-point readings, respectively, can be used to fix the saturated adiabatic curve and the 900-millibar wet-bulb temperature. The wet-bulb potential temperature can be read directly from the saturated adiabatic curve, since each curve is clearly labelled on the current British tephigram. Thus, a 900-millibar wet-bulb potential temperature of 12°C and a 500-millibar dry-bulb temperature of —18°C will give an instability index of 30.

The indices are plotted on a small-scale chart, of say 1 in 10 million, and isopleths are then inserted at one- or two-degree intervals. Index values calculated during the period May–August 1959 ranged from 21 to 35 (Table I). The high values indicated a marked degree of instability and the low values indicated stable conditions.

Evaluation.—This was carried out in two steps. The first step was an attempt to fix the threshold value of the instability index to be associated with the occurrence of significant showers or thunderstorms. Degrees of shower or thunderstorm activity in south-east England were tabulated, together with the value of the instability index calculated from the 2300 GMT temperature sounding at Crawley, Sussex. Reports from 29 meteorological stations within a 75-mile radius of Crawley were studied in order to assess shower or thunder activity for each day during the period May–August 1959. These observations were supplemented by press and radio reports which normally provided confirmation of heavy storms and resultant flooding. Occasions when the activity was directly associated with a frontal discontinuity were excluded, thus eliminating the rather large fluctuations in the value of the indices associated with frontal passage, so making it possible to determine representative values for air-mass instability.

The degree of shower activity was tabulated on a four-point scale. Occasions with showers were divided into two groups, one for showers accompanied by thunderstorms, and the other unaccompanied. There was a further subdivision for days with thunder, in order to categorize occasions with thunderstorms which could be rated as heavy. Similarly, showery days without thunder were split into two categories, by extracting and grouping those days when the showers could only be regarded as very slight and isolated.

The tabulation is produced as Table I. There is a significant shift towards the high indices in the thunderstorm categories, as would be expected. The lowest index value associated with any showers was 25 and on this occasion the precipitation was very slight. The highest value of 35 was associated with heavy thunderstorms, and showers or thunderstorms occurred on all occasions with an instability index of 32 or more, with the probability of 2 to 1 on thunderstorms. It is noteworthy that on the five days with heavy thunderstorms the instability index ranged from 31 to 35, suggesting a threshold value of about 30. The fourteen days when the index stood at 30 are evenly divided, seven with shower activity and seven with none. Nevertheless, thunderstorms occurred on five of the seven showery days.

TABLE I—DEGREES OF SHOWER ACTIVITY AND THE ASSOCIATED INSTABILITY INDEX FOR SOUTH-EAST ENGLAND, MAY–AUGUST 1959

	Instability index (from 2300 GMT Crawley sounding)														
	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35
	number of days														
No showers ..	1	1	0	6	9	5	6	7	9	7	4				
Very slight and isolated showers					1	1	1	3	0	1	1				
Slight–moderate showers								1	1	1	1	3	1		
Showers accompanied by thunderstorms ..									1	5	2	0	2	3	
Heavy thunderstorms											2	0	1	1	1

The assessment covered 89 days. Occasions with frontal activity were excluded.

The second stage of evaluation was carried out using plots of thunderstorm reports and fixes on sources of atmospherics (Sferics). The accuracy of Sferic reports in fixing lightning discharges was adequate for this investigation and has been discussed elsewhere by Horner[3]. A number of maps were produced for days when thunderstorms and Sferic plots were in evidence on the synoptic charts. Once again, occasions with frontal activity were excluded. Instability indices computed from the thirteen soundings made at 2300 GMT were plotted on a small-scale chart and isopleths inserted, together with symbols denoting thunderstorm reports or Sferic fixes extracted from the three-hourly synoptic charts, commencing with the chart for 0001 GMT and terminating with the 2100 GMT chart.

The synoptic chart for 0001 GMT, 12 May 1959, is produced as Figure 1. The Atlantic depression was drifting slowly north-westwards and pressure was intensifying over the British Isles, under the influence of the Scandinavian anticyclone and an intensifying ridge west of Biscay. Figure 2 depicts the instability index chart for 2300 GMT, 11 May 1959. The isopleths delineate an elongated zone of maximum latent instability with the major axis orientated approximately north–south, extending from Normandy to northern Scotland. The zone with high indices thus embraces most of Britain and from the thunderstorm reports and Sferic fixes received on 12 May it is apparent that thundery showers or storms affected large areas of Britain and Normandy.

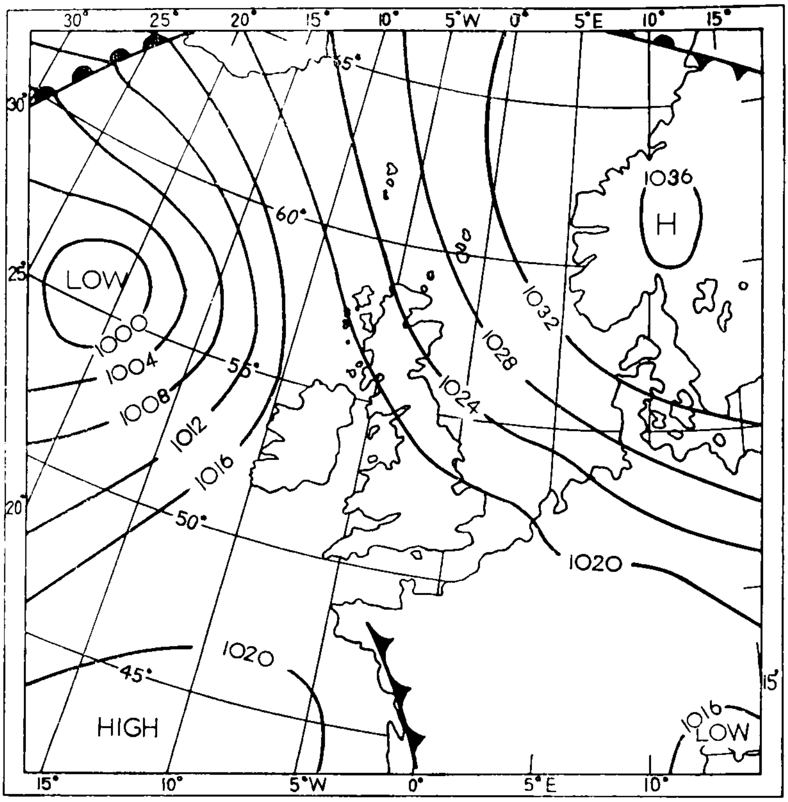

FIGURE 1—SURFACE CHART FOR 0001 GMT, 12 MAY 1959

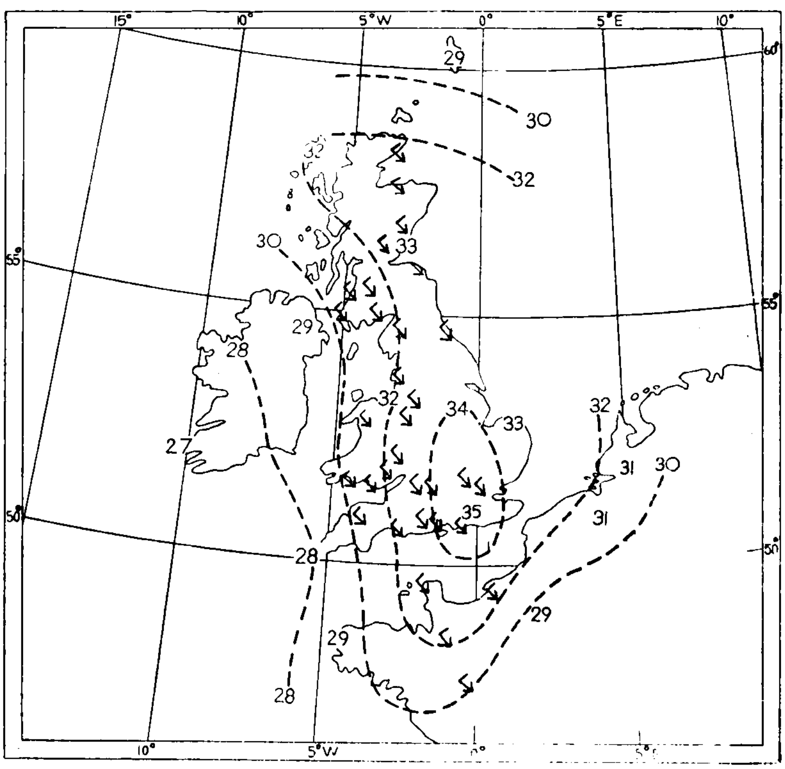

FIGURE 2—INSTABILITY INDEX CHART FOR 2300 GMT, 11 MAY 1959

The broken lines are isopleths of the index, obtained from tephigrams. The symbols denote thunderstorm reports or Sferic fixes for 12 May 1959.

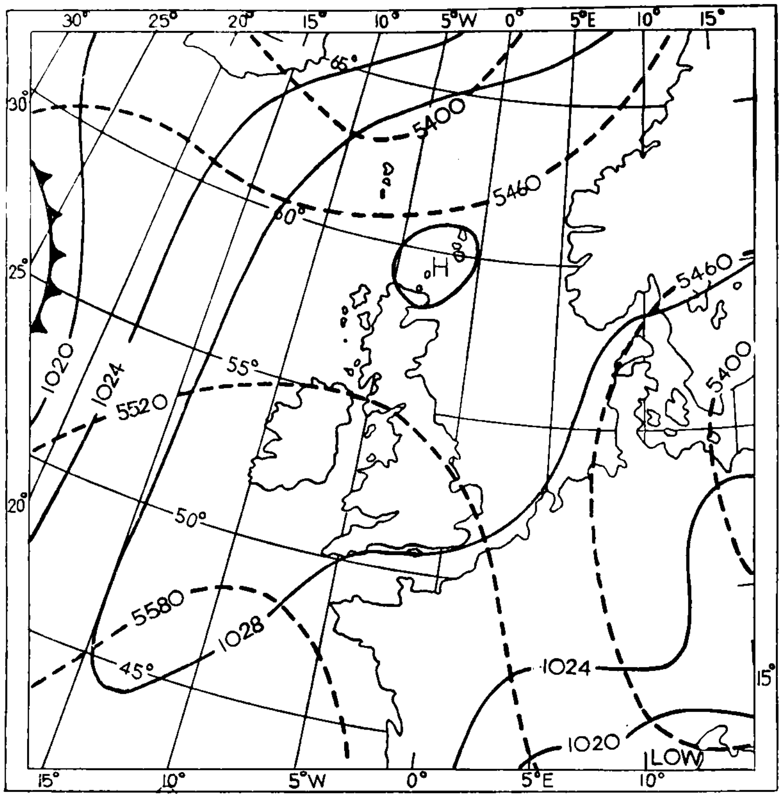

FIGURE 3—SURFACE CHART FOR 0001 GMT, 14 MAY 1959
The broken lines are isopleths of thickness (geopotential metres) for the 1000–500 mb layer.

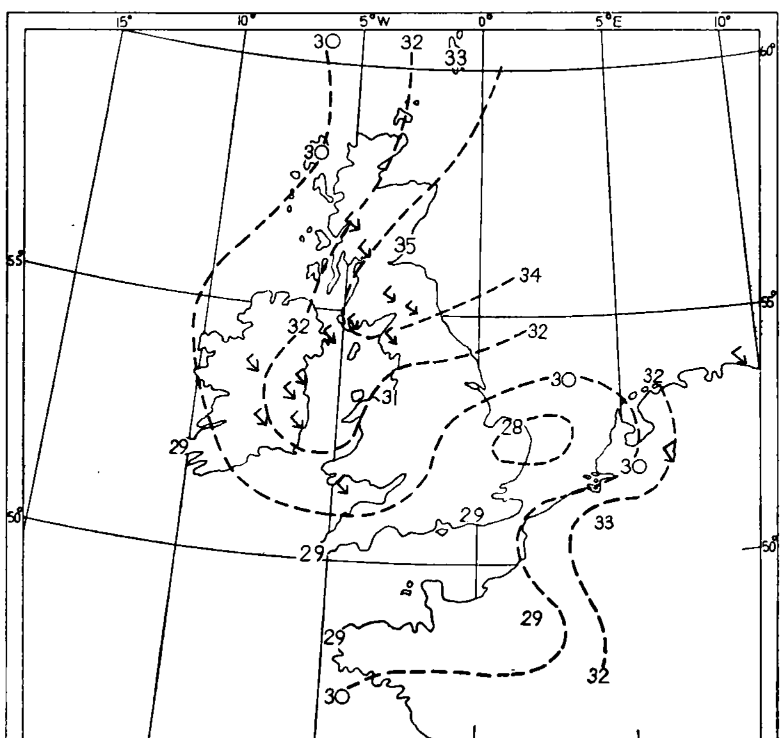

FIGURE 4—INSTABILITY INDEX CHART FOR 2300 GMT, 13 MAY 1959
The broken lines are isopleths of the index, obtained from tephigrams. The symbols denote thunderstorm reports or Sferic fixes for 14 May 1959.

The 24-hour forecast from noon, issued with the *Daily Weather Report* for 12 May 1959 contained the statement, "thunderstorms will occur in places". This is rather vague in the regional sense, but using the instability index as a guide the forecaster could have indicated that thunderstorms would be widespread, but were unlikely to occur in extreme south-west England and Ireland, i.e. in those areas where the index fell below 30 (Figure 2).

On 14 May 1959 an anticyclone centred near Orkney covered the British Isles. The 1000–500-millibar thickness pattern has been superimposed on the synoptic chart (Figure 3); a warm ridge extended across Biscay to the British Isles, with no evidence of cold troughing and associated instability over the region. Nevertheless, the instability index chart for 2300 GMT, 13 May 1959 indicated a zone of instability in the form of a tongue or wedge, with its axis extending from the North Sea, through Berwick and Galloway to south-east Eire, and a more stable region to the south-east, extending from East Anglia to the Dutch coast (Figure 4). Subsequent thunderstorms and Sferic fixes reported during the day were plotted over north-west England, west and south-west Scotland and east and central Eire. There were also reports from the fringe area of the map, in north-west Germany. Aircraft reports also indicated significant cumulonimbus development during the day. An aircraft over the Isle of Man at 0800 GMT reported an isolated cumulonimbus top at 17,000 feet but by 1315 GMT cumulonimbus tops ranging from 30,000 to 35,000 feet were reported in a position about 25 miles north of Silloth, Cumberland. Figure 4 indicates good agreement between the fixes and the isopleth pattern and an inspection of this chart, and the previous chart for 11 May 1959, suggests a critical or threshold value of about 30, for the instability index, in agreement with the inference drawn from Table I.

The 2300 GMT soundings for Camborne, Aughton (Liverpool) and Aldergrove of 13 May 1959, indicated rather moist air aloft, and the forecast issued with the *Daily Weather Report* for 14 May 1959 included the statement, "scattered thunderstorms may break out in western areas later this afternoon". The forecaster inspecting the instability index chart and noting the prominent tongue or wedge pattern, might have been prompted to state that thunderstorms were most likely to occur in northern England, southern Scotland and eastern Ireland. This would have been preferable to the rather vague reference to "western areas". The degree of confidence or probability could also have been raised, assuming a threshold index value of about 30 and noting that the plotted indices ranged as high as 34 to 35.

Figure 6 depicts the instability index chart for 2300 GMT, 27 July 1959, and the isopleths indicate a high degree of instability over southern and eastern regions of the British Isles. The assumption of thundery activity, with indices well above the 30 mark, was borne out by the numerous Sferic fixes and thunderstorm reports plotted during the succeeding day; these extended from the Low Countries and North Sea across England to eastern Ireland. The synoptic chart for 0001 GMT, 28 July 1959 (Figure 5) indicated a slow-moving depression centred near the Solway Firth and an associated cold front extending from the North Sea, through the German Bight to south-west France.

The 24-hour forecast from noon, issued with the *Daily Weather Report* for 28 July 1959, merely stated that "thunderstorms are likely in places". On the

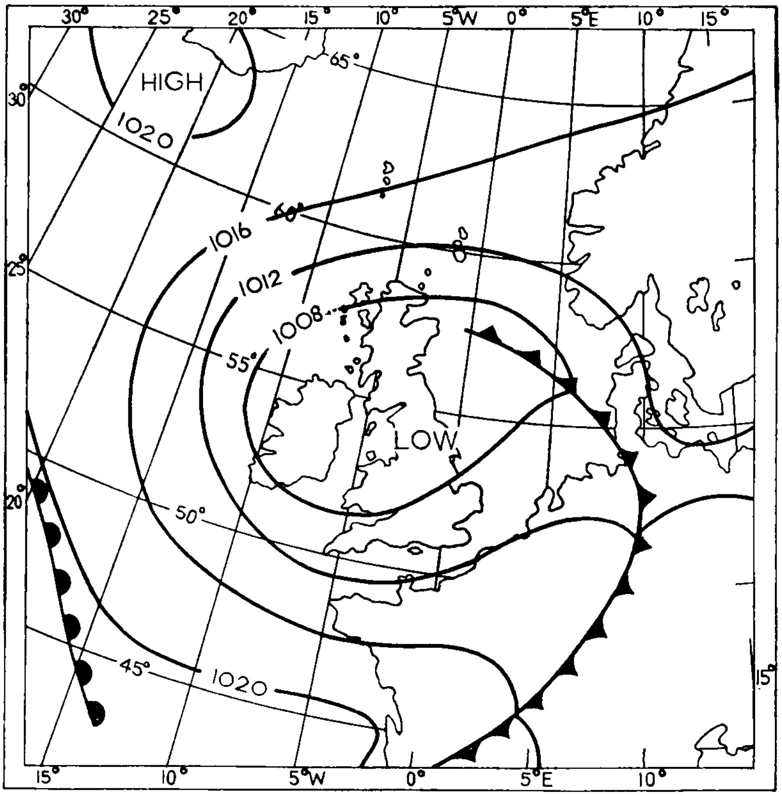

FIGURE 5—SURFACE CHART FOR 0001 GMT, 28 JULY 1959

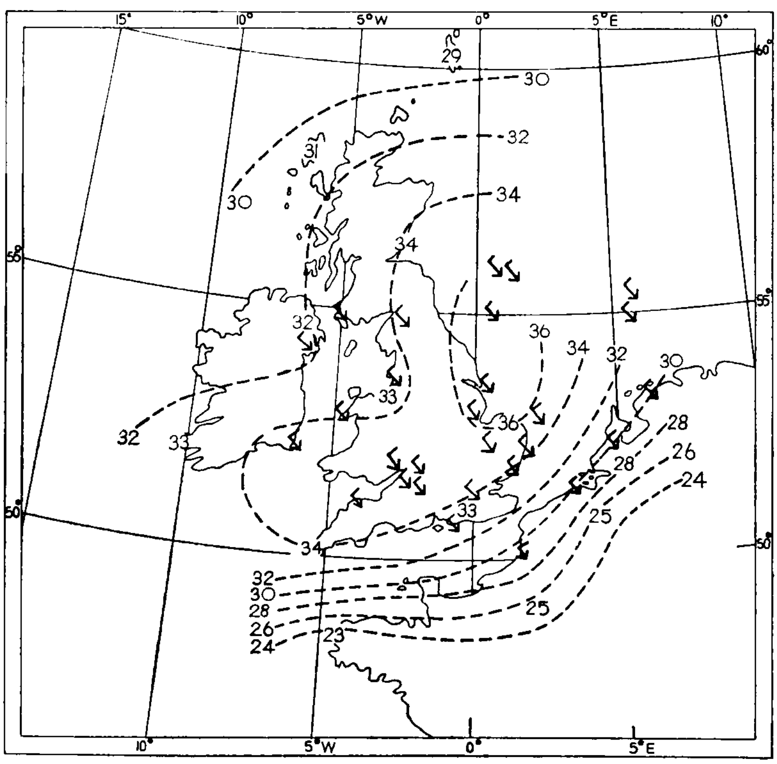

FIGURE 6—INSTABILITY INDEX CHART FOR 2300 GMT, 27 JULY 1959
The broken lines are isopleths of the index, obtained from tephigrams. The symbols denote thunderstorm reports or Sferic fixes for 28 July 1959.

following day there was little change in the synoptic situation and the instability index over most of Britain exceeded 32, indicating a strong probability of further widespread thunderstorm activity. Nevertheless, the forecast issued with the *Daily Weather Report* for 29 July 1959 did not mention thunderstorms although, in fact, widespread thunderstorms did occur. The forecaster, using the instability index chart as an additional aid, could have confidently predicted widespread thunderstorms, or at least indicated a strong probability of widespread storms occurring on both 28 and 29 July 1959.

Conclusions.—It is suggested that a chart based on the instability index would assist the regional forecaster, particularly during the summer thunderstorm régime. The simple computations and objective method mean that charts can be prepared very rapidly from the 2300 GMT radiosonde data; thus the deduced information is available for the early forecast bulletins.

In non-frontal situations, an index value exceeding 30 should alert the forecaster to the prospect of significant showers accompanied by thunderstorms.

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SYNOPTIC FACTORS ASSOCIATED WITH RELAXING THERMAL TROUGHS AND THEIR PREDICTION VALUE

By M. K. MILES and G. A. WATT

Summary.—All troughs in the 1000–500-millibar thickness lines which underwent a certain minimum amount of relaxation between the east coast of America and about 15°E in the years 1953–59 (inclusive) form the basis of the study. It is found that, despite the strong contribution of convective warming over the west and central Atlantic, the process is essentially a dynamical one. The two main agents in this appear to be:

(i) an upwind trough coming, or forming, within 35°–40° longitude and

(ii) an anticyclone (usually centred north of 40° N) within 35°–40° longitude downwind.

They are usually both present when relaxation starts. A forecasting rule based on the sum of these two spacings was tested on data for 15 months 1960–61, and 82 per cent of the forecasts made were correct.

Relaxation has the following synoptic consequences:

(i) a rise in surface pressure on the associated cold front, together with a reduction in rainfall intensity and suppression of wave formation;

(ii) an increase in static stability within the thermal trough.

Introduction.—Since charts of thickness (1000–500 millibars) became a working tool just after the end of World War II, the warming (and weakening) of cold areas (represented by troughs in the thickness lines and usually known as thermal troughs) has been recognized as an important occurrence. The process, involving northward movement of the thickness lines in the trough, early came to be described as relaxation: the term is now so securely built into the synoptic meteorologist's vocabulary that it seems expedient to continue to use it.

Figures 1 to 4 illustrate a case in which four thickness lines of a trough in the east Atlantic relaxed although the warmest line did not. It is not uncommon for this to happen and in such cases there is either a surface low or a cut-off cold pool associated with the southern part of the trough.

This study was designed to discover, more precisely than is known at present, the synoptic factors which favour and accompany the process of relaxation.

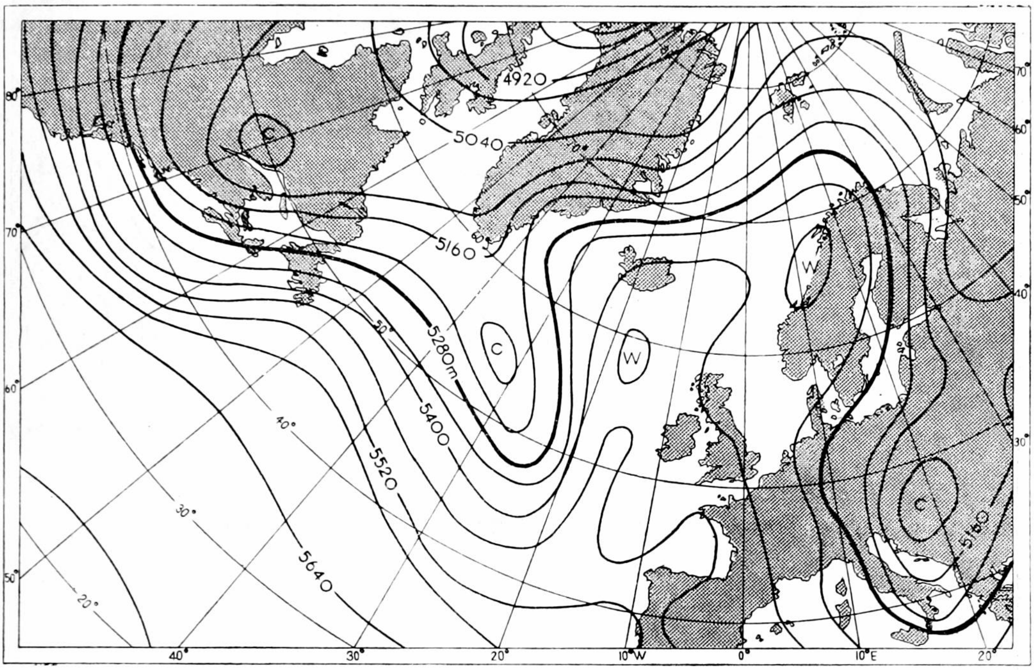

FIGURE 1—1000–500 MB THICKNESS CHART FOR 1500 GMT, 9 MARCH 1956

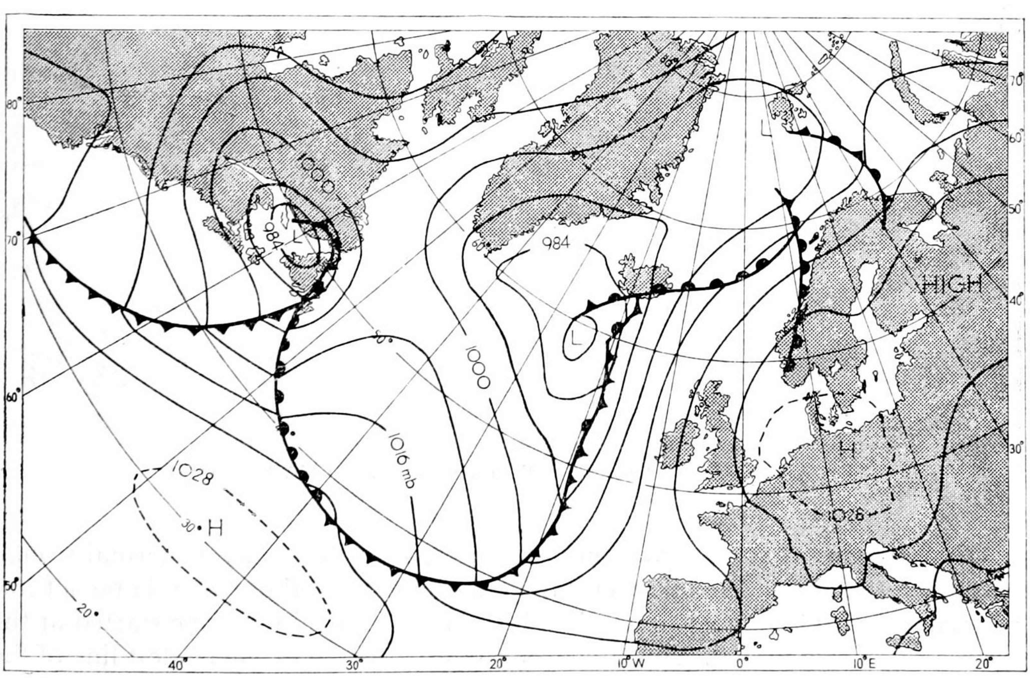

FIGURE 2—SURFACE CHART FOR 1500 GMT, 9 MARCH 1956

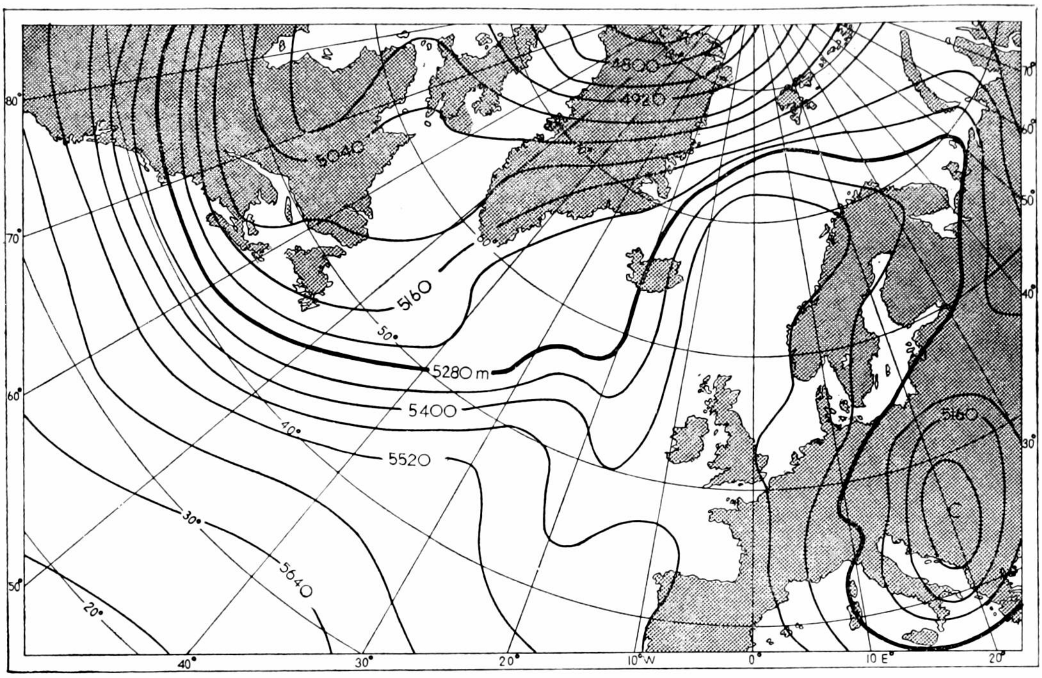

FIGURE 3—1000–500 MB THICKNESS CHART FOR 1500 GMT, 10 MARCH 1956

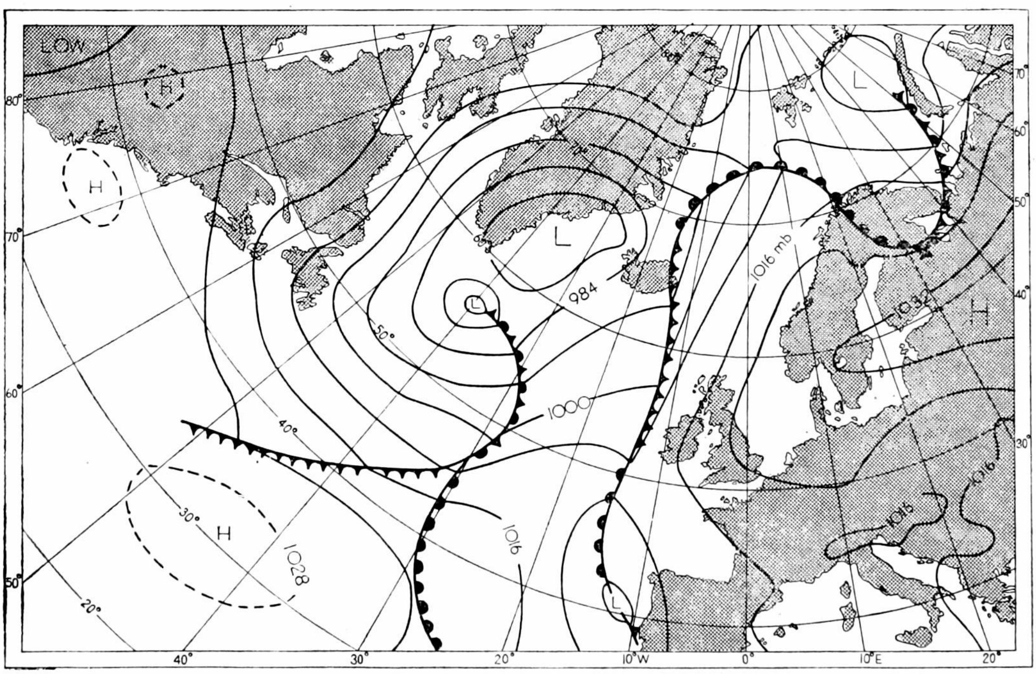

FIGURE 4—SURFACE CHART FOR 1500 GMT, 10 MARCH 1956

Observational material.—For the purpose of this study a thermal trough was taken to have relaxed if several thickness lines moved north and one of these at least 5° latitude in a 24-hour period. Relaxation is said to have started at the beginning of this period. This line was usually other than the coldest line of the trough, and will henceforth be referred to as the *defining thickness line* of the trough.

All thermal troughs which relaxed between the American coast and 15°E were noted for the seven years 1953 to 1959. Calling the time when relaxation starts *t*, the following data were extracted for *t*—24 hours, *t* and *t* + 24 hours for each case:

(i) the latitude and longitude of the most southerly point of the defining thickness line of the trough;

(ii) the latitude and longitude of the most southerly point of this defining thickness line in the next upstream thermal trough;

(iii) the latitudes and longitudes of the most northerly point of this defining thickness line in the adjoining thermal ridges;

(iv) the latitude, longitude and central pressure of the nearest downwind anticyclone, provided there was not another thermal trough in between;

(v) the latitude and longitude and central pressure of the nearest upwind anticyclone, provided this was situated east of the upwind thermal ridge.

Synoptic statistics associated with relaxation

Characteristics of the relaxing trough.—During the first 24 hours of relaxation the average speed of the 277 relaxing troughs was 13° longitude per day (standard deviation 6.3) and the average northward movement of the defining thickness line was 7° latitude. Of these, 162 could be identified at *t* + 48 and the average speed was 14° longitude per day from *t* + 24 to *t* + 48 and the average relaxation was 6° latitude. The average movement over the 24 hours preceding relaxation was 12° longitude and the average meridional movement of the defining thickness line was 0° latitude.

The latitude and average value of the defining thickness line is shown in Table I for each month of the year. It appears that relaxation tends to occur most frequently in a fairly restricted latitude band between the subtropical

TABLE I—LATITUDE AND AVERAGE VALUE OF DEFINING THICKNESS LINE

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Average latitude (°N)	40	40	41	41	43	45	48	46	47	43	43	42
Standard deviation	4.1	4.3	3.1	4.3	4.9	3.7	6.8	3.2	4.2	4.6	4.9	5.1
Average value of defining thickness line (decametres)	536	535	534	536	545	548	554	553	545	545	539	535
No. of cases	26	37	25	25	37	17	15	18	19	17	23	19

high-pressure belt (30°–35°N) and the mean pressure trough at about 60°N. The seasonal movement is of the same order as the seasonal shift of these mean circulation features.

Relaxation had begun on the majority of occasions by the time the trough reached 30°W. This is especially so in winter (November to March inclusive) when the proportion is 85 per cent. In the months June to August (inclusive) the percentage of troughs relaxing east of 30°W rises to 58 per cent. The greater non-adiabatic warming of cold continental air moving into the west Atlantic in winter is clearly reflected in these figures.

The mean thickness gradient was measured over a distance of 400 nautical miles at a representative position ahead of and behind the trough at *t*. This allowed a diffluence ratio (gradient behind to gradient ahead) to be worked out for each trough. If ratios equal to or greater than 1.5 are taken to represent markedly diffluent troughs, then less than 4 per cent of the total were of this kind at the start of relaxation. About 40 per cent were confluent if this is defined by values of the ratio equal to or less than 0.7. The remaining 56 per cent cannot be described as markedly confluent or diffluent. Thus it may be concluded that though confluence is commonly associated with relaxation it is by no means a necessary condition for it.

Relation to upwind features

(*a*) Upwind thermal trough.—It has been the common experience of synoptic meteorologists that the formation of a new thickness trough upwind was, under some circumstances, associated with the relaxation of its downwind neighbour. This study has shown that one of the circumstances is the spacing between the two. Table II shows the mean spacing at $t - 24$, t and $t + 24$, and the standard deviation of these quantities. The value is significantly shorter than the more usual spacing of about 60° longitude* between troughs in the westerlies, even before relaxation begins and it gets less as the process goes on. Although there is a certain amount of scatter in the spacing distribution, 77 per cent of the relaxations began when the upwind trough was between 30° and 49° longitude away.

TABLE II—MEAN SPACING FROM UPWIND THERMAL TROUGH BEFORE AND DURING RELAXATION OF A THERMAL TROUGH

	$t - 24$	t	$t + 24$
Spacing (° longitude)	42	38	34
Standard deviation	9.2	8.5	10.3
No. of cases	184	263	270

The upwind thermal trough underwent some meridional extension between $t - 24$ and t in just over half of the cases. On some occasions it first appeared as a recognizable feature in this interval, and was frequently only a small-amplitude trough at time *t*. Its average movement was 14° longitude from $t - 24$ to t, and 16° longitude in the next 24 hours.

(*b*) Upwind thermal ridge.—Table III shows the relation of this feature to the relaxing and upwind thermal troughs at three times during the process. It is evident that while the short spacing between the upwind thermal trough and the ridge remains sensibly constant, the thermal ridge is steadily approaching the relaxing thermal trough. This finding confirms the result given by Miles[1]

TABLE III—SPACING OF UPWIND THERMAL RIDGE FROM RELAXING AND UPWIND THERMAL TROUGHS

	$t - 24$	t	$t + 24$
Spacing from relaxing trough (° longitude)	25	21	15
Standard deviation	6.8	5.6	5.4
Spacing from upwind trough (° longitude)	17	17	18
Standard deviation	6.7	6.1	7.9

*The mean value for 26 trough pairs used in a test of the Rossby formula at the Central Forecasting Office was 58° longitude.

based on a study of one year's troughs, though it indicates that the relaxation actually begins when the thermal ridge is rather farther away than the figure of 15° longitude that he found.

The average movement of the thermal ridge was 16° longitude between $t - 24$ and t and 14° longitude in the next 24 hours, i.e. very nearly the same as the upwind thermal trough. These two features represent a fairly mobile perturbation of small to moderate amplitude. The surface depression associated with this thermal pattern deepened by an average of 11 millibars between $t - 24$ and $t + 24$ and moved at an average speed of 16° longitude per day over the same interval. It very rarely moved round the crest of the upwind thermal ridge.

(*c*) Upwind surface anticyclone.—As Table IV shows this anticyclone is rather close to the relaxing trough.

TABLE IV—SPACING BETWEEN UPWIND ANTICYCLONE AND RELAXING TROUGH

	$t - 24$	t	$t + 24$
Spacing (° longitude)	13	12	9
Standard deviation	5.9	5.2	4.9

By way of comparison it was found that the mean spacing of 98 thermal troughs from the nearest upwind surface anticyclone after 24 hours of meridional extension was 22° longitude.

Relation to downwind features

(*a*) Downwind thermal ridge.—The data in Table V indicate that the downwind thermal ridge is already a fairly large-amplitude feature 24 hours before relaxation begins and grows on average a further 5° latitude in this interval. It moves east at the same speed as the relaxing trough. It is noteworthy that although the amplitude decreases between t and $t + 24$, the defining thickness line on the crest moves north by an average amount of 3° latitude.

TABLE V—SPACING AND AMPLITUDE DATA FOR RELAXING TROUGH AND DOWNWIND THERMAL RIDGE

	$t - 24$	t	$t + 24$
Amplitude * (° latitude)	15	20	16
Standard deviation	7.2	5.9	5.9
Spacing (° longitude)	21	21	20
Standard deviation	8.8	10.7	10.4

*defined as latitudinal difference of defining thickness line in relaxing trough and downwind ridge

(*b*) Downwind anticyclone.—There was a downwind anticyclone associated with about 90 per cent of the relaxing troughs. Before and during the relaxation, as the figures in Table VI show, the relaxing trough was getting nearer to the centre of this anticyclone. Relaxation began at a rather wide variety of distances from the anticyclone. The modal value lay between 30° and 35° longitude but there were a substantial number of cases when the value was more than 45° longitude, especially when the anticyclone was over north-west Europe.

TABLE VI—AVERAGE SPACING BETWEEN RELAXING TROUGH AND DOWNWIND ANTICYCLONE

	$t - 24$	t	$t + 24$
Spacing (° longitude)	43	37	29
Standard deviation	13.4	11.3	11.9
No. of cases	243	251	246

However, it was very rare for it to exceed 45° longitude when the defining thickness line of the trough was less than 5° latitude south of the centre of the anticyclone. The centre of the anticyclone was on average 9° latitude north of the defining thickness line in the trough at the start of relaxation: this was reduced to 2° latitude at $t + 24$.

TABLE VII—AVERAGE LATITUDE OF DOWNWIND ANTICYCLONE AT START OF TROUGH RELAXATION

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Average latitude (°N)	47	47	50	49	53	64	57	53	52	51	49	49

The average latitude of all the downwind anticyclones at time t was 51°N (standard deviation 8.7° latitude), indicating that a large preponderance of them were well north of the subtropical high-pressure belt. In fact 90 per cent of them were north of 40°N. From Table VII it is evident that many of the relaxations in June and July and to a lesser extent in May are associated with fairly high-latitude blocking anticyclones.

There is some evidence that the latitude of the downwind anticyclone determines whether the trough accelerates or decelerates during relaxation. If acceleration is defined as a change of 5° longitude or more in successive 24-hourly movements centred around the start of relaxation (t), then 78 troughs accelerated and 69 retarded. The mean latitude of the downwind anticyclone for each of these two classes was 48° N and 54° N respectively and the standard deviations of these two means were 0.9 and 1.2. This indicates that the difference of 6° in the two means is probably significant, and that acceleration is more likely to accompany relaxation when the downwind anticyclone is south of 50° N.

Spacing of the anticyclones adjacent to the relaxing trough.—At t the mean spacing between the upwind and downwind anticyclones was 48° longitude. When the downwind anticyclone was centred west of 10°E (over 75 per cent of the cases) 70 per cent of the spacings lay between 35° and 55° longitude (inclusive). For the remaining cases, spacings greater than 60° longitude were about as frequent as those less than 60° longitude. In this connexion it is interesting to note that Miles and Leaf[2] found evidence that a trough does not readily extend between two anticyclones less than 60° longitude apart. It is also significant that the spacing of the upwind and downwind anticyclones decreased at an average rate of 9° longitude per day from $t - 24$ to $t + 24$.

Some synoptic occurrences associated with relaxation

Behaviour of associated surface depressions.—The latitude of the surface depressions associated with the relaxing trough was found to increase on average 5° latitude per day from $t - 24$ to $t + 24$. The average eastward movement was only a little over half that of the relaxing trough. By contrast the surface depression associated with the upwind ridge–trough pattern moved much faster and the two systems were coming together at an average rate of 8°–9° longitude per day.

Surface pressure changes.—There was usually a rise of surface pressure over the central region of the relaxing trough. Although the statistics given earlier show that the centre of the anticyclone behind the relaxing trough did not move north during relaxation, it was not unusual for a ridge from it to extend north-eastwards with the relaxing trough.

There was nearly always a rise of surface pressure on that part of the associated cold front between the latitude of the defining thickness line of the trough at t and $t + 24$. For 17 cold fronts associated with relaxing troughs which were east of 20° W at $t + 24$, the average rise was 6 millibars for points at the same latitude on the front.

Changes in frontal activity.—The behaviour of the rain belt with 38 cold fronts crossing the British Isles and north-west Europe was examined for the period t to $t + 24$. For 32 of them the amount of rain was small or, if it was moderate at t, had become slight or ceased by $t + 24$.

Wave formation on the associated cold front.—A sample of 77 cold fronts associated with relaxing troughs which were east of 20° W at $t + 24$, was examined for new wave formation in the period t to $t + 24$. On only two* of them was there wave formation after t. At time t the average length of the baroclinic zone ahead of the trough was about 1200 nautical miles and it decreased during relaxation. Since there was almost always sufficient thermal gradient ahead of the trough, this result supports the conclusion reached by Sawyer[3] that spacing is the dominant factor in determining wave formation on cold fronts.

The mechanism of trough relaxation.—A detailed study of several individual cases showed that the relaxation was partly due to non-adiabatic warming of the layer from the surface to about 700 millibars and partly to dynamically produced effects most apparent above this level. For example, there was usually evidence of subsided cold air between 700 and about 550 millibars and sometimes of warmer moister air above this. Trajectories at 500 millibars indicated that this was air carried into the trough from the upwind thermal ridge. The contribution of non-adiabatic warming becomes less in the east Atlantic; in this region in summer, the effect is almost entirely due to adiabatic warming.

This dynamical effect appears to be mainly due to:

(i) the short spacing between the relaxing trough and the upwind trough and

(ii) the approach to an anticyclone strong enough to have a moderate-to-large-amplitude thermal ridge 15° to 20° longitude to the west of it on average.

The increasing nearness of the upwind thermal ridge as relaxation proceeds e.g. 25°, 21° and 15° longitude at $t - 24$, t and $t + 24$ respectively, appears to be an important upwind factor. A few cases of relaxation were encountered in which the upwind thermal ridge was within 20° longitude but there was no downwind anticyclone and the distance to the upwind trough was more than two standard deviations greater than the mean value associated with relaxation. This corresponds to an asymmetric pattern with a long belt of south-westerly flow between the upwind trough and the ridge: it is usually associated with a wide, open warm sector i.e. minimum width 25° longitude. Asymmetry in the other direction, however, i.e. upwind thermal ridge 35° from the trough and 15° from the upwind trough appears not to be associated with relaxation. Indeed relaxation only occurred when the thermal ridge was between 30° and 35° away if its amplitude was less than 15° latitude.

*One of these was with an unusually large and intense trough (1000–500 -millibar thickness anomaly at 51°N 45°W at t was —36 decametres) on 28 January 1957.

However, most cases of relaxation occurred when both conditions were satisfied, i.e. the closer than usual upwind trough and the presence some 30°–50° longitude downwind of a well developed anticyclone. The intensity of these anticyclones is indicated by the size of the thermal ridge associated with them, and their central pressures which averaged 1032 millibars between October and January and 1028 millibars from February to September. Only two of them had central pressures less than 1020 millibars.

A possible prediction parameter.—It appears that the two principal factors determining relaxation may be represented by requiring that the distance between the downwind anticyclone and the upwind thermal trough should fall below some particular value. The individual distances are shown as the ordinate of the scatter diagram in Figure 5 with the longitude of the downwind anticyclone as the abscissa. There appears to be a fairly well defined upper

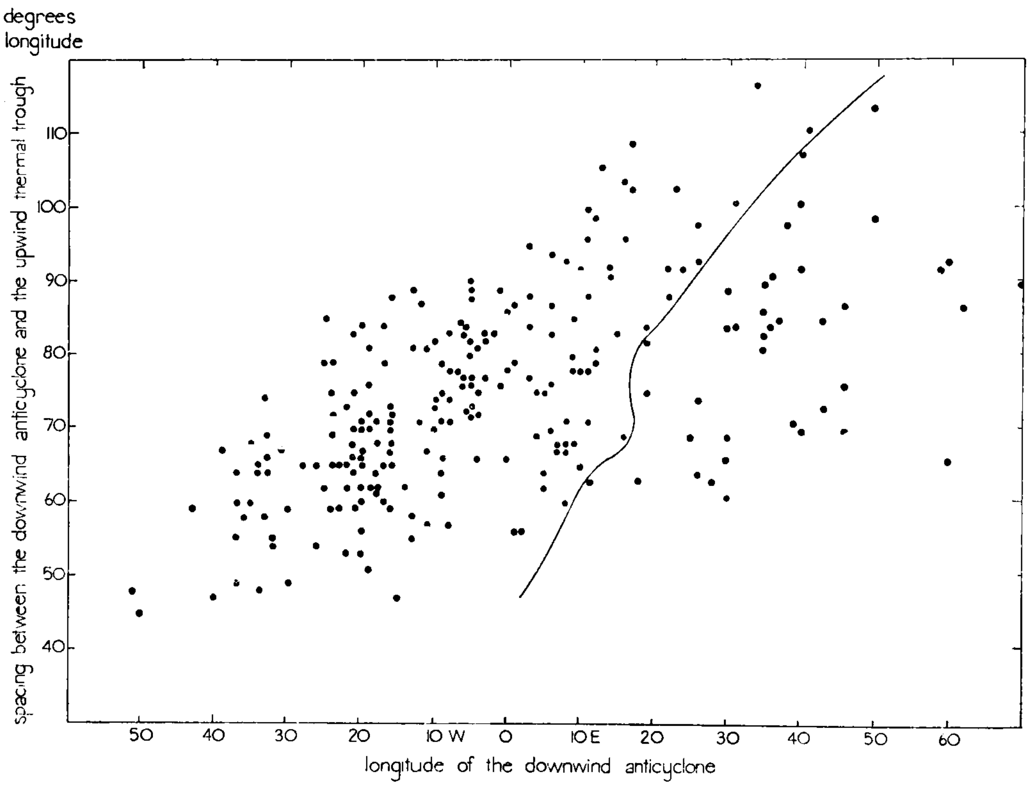

FIGURE 5—RELATION BETWEEN SPACING AND LONGITUDE OF DOWNWIND ANTICYCLONE AT START OF RELAXATION

All cases to the right of the line occurred in the months March–September.

limit which increases the farther east the relaxation occurs. As there is also an effect due to the latitude of the downwind anticyclone it was necessary to examine the values geographically. The result is shown in Figure 6 in which areas are delineated with a representative mean value. A value is also shown, above which 25 per cent or less of the relaxations occurred.

A preliminary test on independent data showed that the use of this value eliminated obvious over-forecasting of relaxation. It may also be expected to

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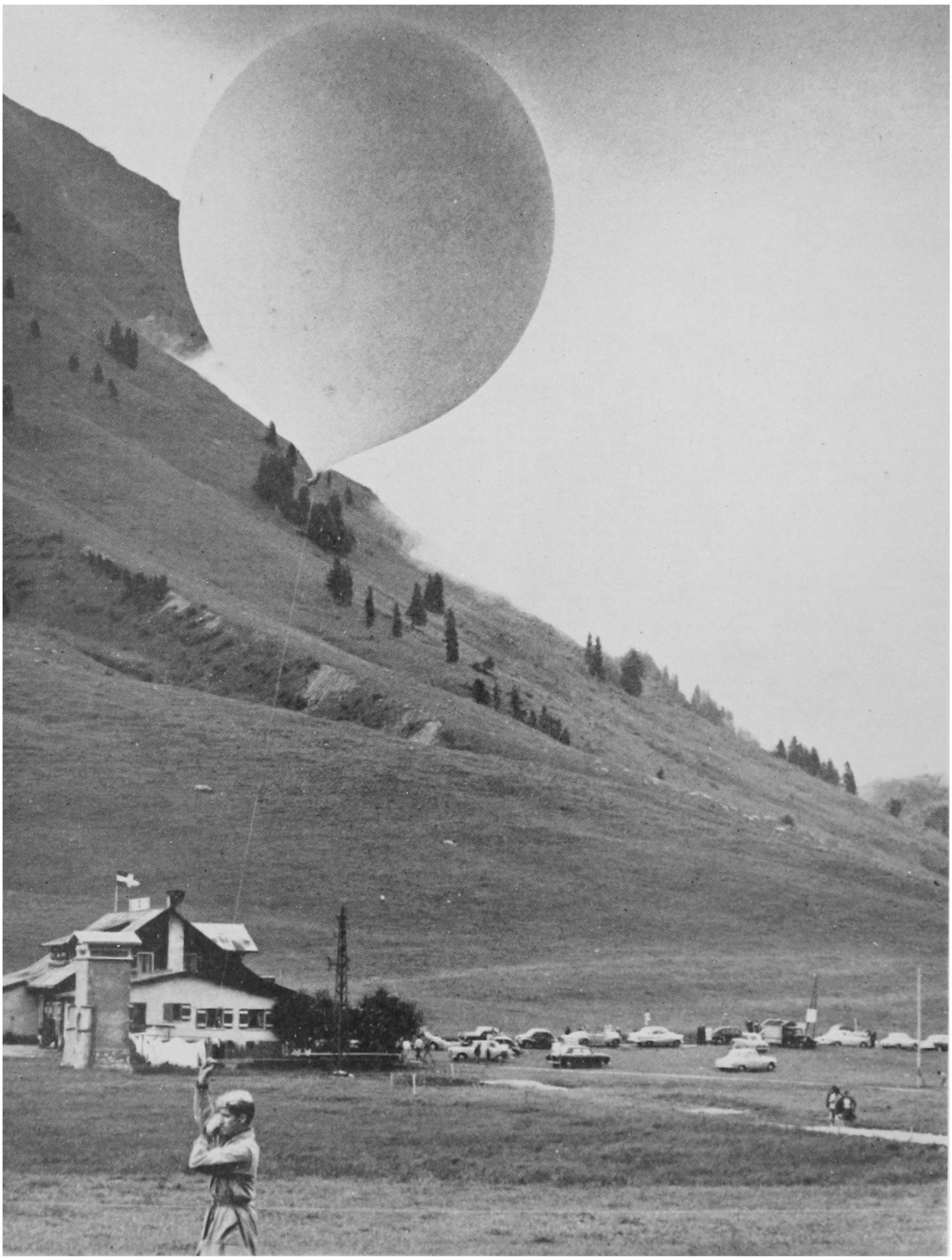
Photograph by G. A. Tunnell.

"WEATHER REPORTER" (IN FOREGROUND) AND "WEATHER ADVISER" IN THE JAMES WATT DOCK AT GREENOCK



Photograph by R. S. Scorer

RADIOSONDE BALLOON BEFORE RELEASE

(see p. 140)

Photograph by R. S. Scorer

RADIOSONDE BALLOON AFTER RELEASE

(see p. 140)

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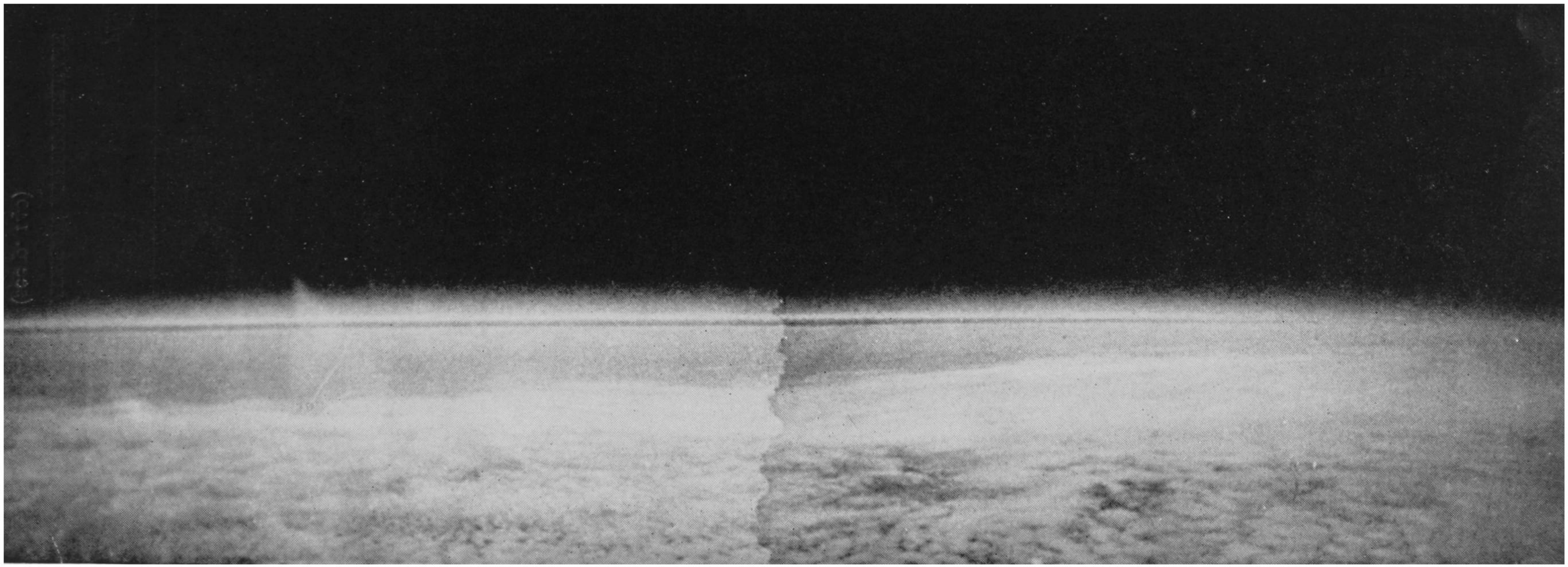

Photograph by CSIRO.

DUST LAYER ABOVE TROPOPAUSE PHOTOGRAPHED FROM A HEIGHT OF 66,000 FEET

(see p. 139)

lead to failure to forecast at least 25 per cent of relaxations and a further small percentage which occur either without an upwind thermal trough or a downwind anticyclone.

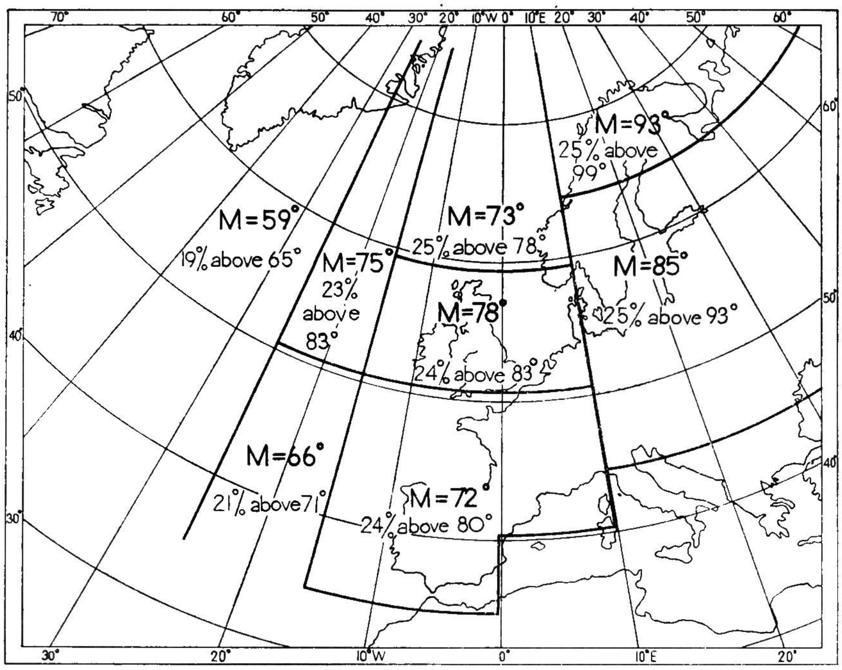

FIGURE 6—MEAN VALUE (M) OF THE SPACING BETWEEN DOWNWIND ANTICYCLONE AND UPWIND THERMAL TROUGH IN RELATION TO THE AREA IN WHICH THE ANTICYCLONE IS CENTRED

A proposed forecasting rule and its test.—Whenever there is both an upwind thermal trough and a downwind anticyclone and the spacing between them does not exceed the value appropriate to the region in which the anticyclone is centred (see Figure 7), relaxation of the trough by more than 5° latitude in 24 hours should be expected provided the following further conditions are all satisfied:

(i) the amplitude of the downwind thermal ridge is greater than 5° latitude, but if less than 15° latitude it shall have increased by at least 2° latitude in the preceding 24 hours,

(ii) the central pressure of the downwind anticyclone is at least 1020 millibars for the months February to September (inclusive) and at least 1024 millibars for the months October to January,

(iii) the upwind thermal ridge is not more than 35° longitude away from the trough, and, if it is between 30° and 35° longitude away, the amplitude is less than 15° latitude.

It should be borne in mind that the upwind thermal trough is not necessarily the next large-amplitude trough upwind. A quite small-amplitude feature (even less than 5° latitude) should be used especially if it is in a region of northerly surface flow.

The test of this rule was carried out on all well defined troughs between the American coast and 10°E occurring in the period January 1960 to March 1961 inclusive. A forecast was made on every day with a suitable trough, provided relaxation had not already begun. A forecast of relaxation was considered to be correct if at least one thickness line of the trough (other than the coldest) moved north 5° latitude or more in a 24-hour period starting either at the time of forecast or 12 hours later. 201 forecasts were made and Table VIII is a contingency table showing the four combinations of forecast and outcome.

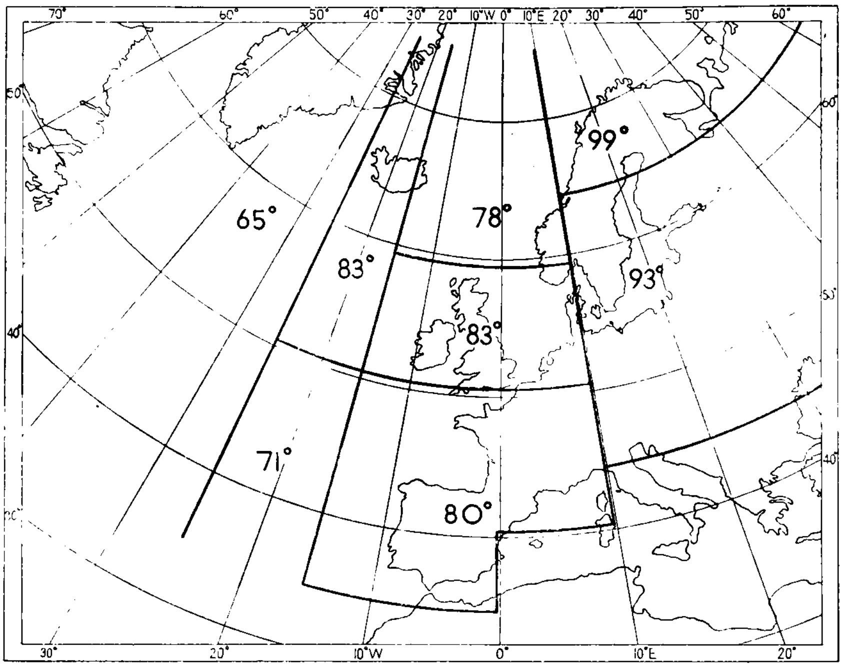

FIGURE 7—MAXIMUM VALUE BETWEEN DOWNWIND ANTICYCLONE AND UPWIND THERMAL TROUGH FOR EXPECTATION OF RELAXATION

TABLE VIII—TABLE OF RELAXATION FORECASTS AND OUTCOMES

		Forecast: Relaxation	Forecast: No relaxation
Outcome	Relaxation	32	19
	No relaxation	18	132

This shows that the rule as it stands does not seriously over-forecast relaxation and that when no relaxation is indicated a considerable degree of skill is obtained. The fact that 37 per cent of cases of relaxation were not forecast is not surprising. The fact that 36 per cent of the forecasts of relaxation were not borne out calls for some comment. A few of these were cases when less than 5° of relaxation occurred and there was one at least in which the upwind depression had filled, in the 12 hours before the forecast was made. The other situations in which expected relaxation did not occur may be summarized under two headings:

(i) very meridional situations when short spacings and large amplitudes are typical,

(ii) troughs occurring in very close association with large slow-moving surface depressions.

Conclusions.—Relaxation of thermal troughs is a fairly common occurrence in the west and central Atlantic in the winter half of the year. It is much less common in the east Atlantic and the north-west European seaboard, though in the months June to August (inclusive) occurrences east of 20°W slightly exceed those west of this longitude.

Over the west and central Atlantic convective warming from the sea plays an important part, though dynamical warming also occurs and is mainly apparent above the 700-millibar level. Over the east Atlantic and the British Isles dynamical warming is the main factor, and in summer probably the only factor. This dynamical process appears to be mainly attributable to a combination of the following two conditions:

(i) a thermal trough approaching (or forming) within 35°–40° longitude upwind of the relaxing one,

(ii) the relaxing trough approaching within 35°–40° longitude of a well developed anticyclone centred north of the subtropical high-pressure belt (average latitude 51°N).

Some important consequences of trough relaxation over the east Atlantic and the British Isles are:

(*a*) a rise of pressure along the associated cold front,

(*b*) a reduction in the intensity of rain associated with the cold front,

(*c*) almost complete suppression of wave formation on the cold front, and

(*d*) an increase in the static stability of the air in the thermal trough.

There is usually a surface anticyclone situated rather close behind the axis of a relaxing trough. The average separation of 12° longitude is significantly less than the average 22° longitude for a sample of extending troughs.

The spacing between the upwind thermal trough and the downwind anticyclone has been used as the basis for a rule for predicting relaxation. In a test on 15 months' data in 1960–61 a successful forecast of relaxation or no relaxation was obtained on 82 per cent of occasions.

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2. MILES, M. K. and LEAF, G. G.; The occurrence and prediction of cold northerly-type spells over the British Isles in winter. *Met. Mag., London*, **90**, 1961, p. 227.
3. London, Meteorological Office; Meteorological Office discussion. *Met. Mag., London*, **79**, 1950, p. 146.

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A NOTE ON SEVERE TURBULENCE AT RENFREW (GLASGOW) BETWEEN 500 AND 1000 FEET ON 26 JANUARY 1961

By R. WILSON

Synoptic situation.—At 1800 GMT, 26 January 1961 (see Figure 1), an intense depression about 150 miles south of Iceland was moving slowly north-north-west and the associated fronts were affecting western and northern areas of the British Isles. Pressure was high over Europe. The main cold front curved south-westwards to an active wave depression some 550 miles west-south-west of Valentia, the wave deepening and moving very rapidly north-eastwards.

A warm occlusion, which lay from Cape Wrath to Rothesay to Bangor thence southwards, was moving steadily eastwards and passed Renfrew around 2100 GMT. A warm front lying north-west to south-east over the western half of Northern Ireland was moving steadily eastwards; by midnight it was lying

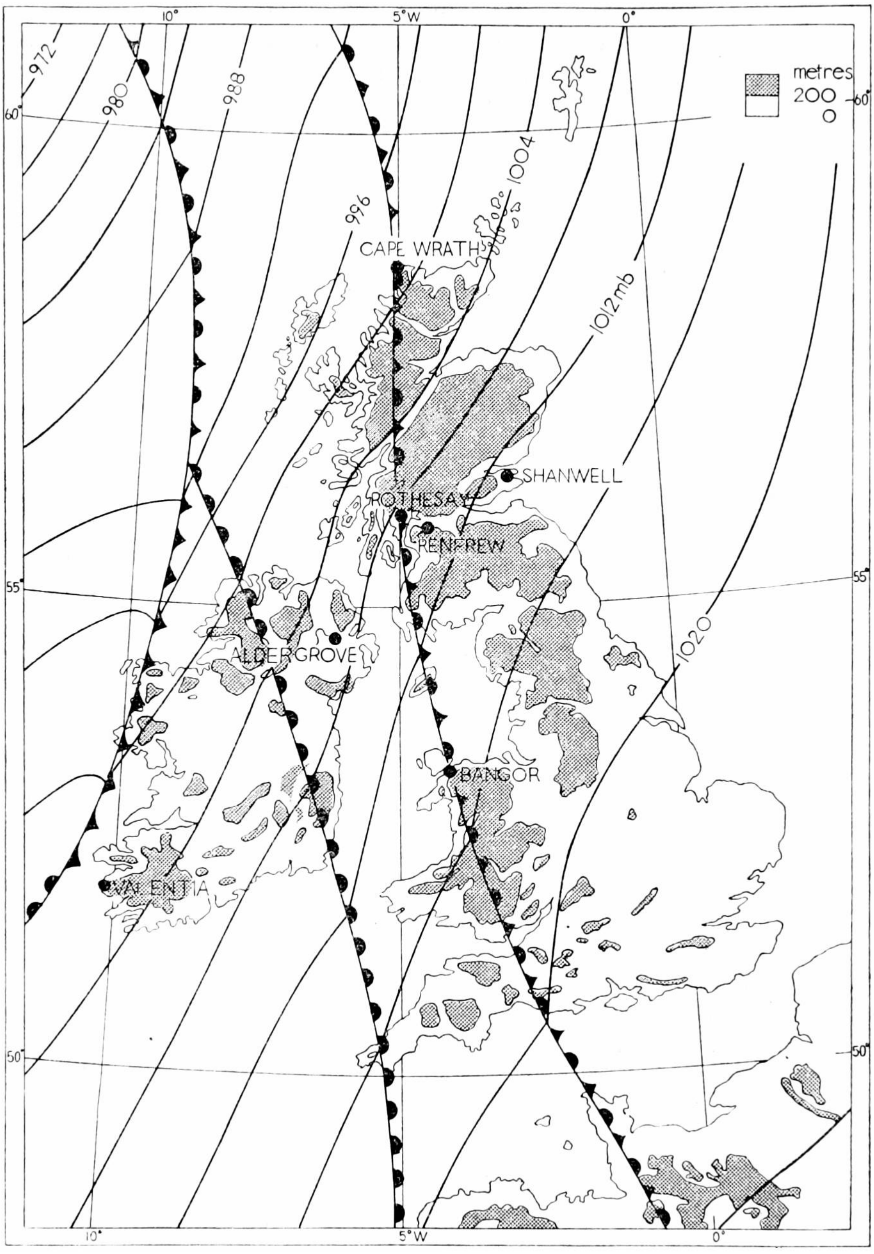

FIGURE 1—SURFACE CHART FOR 1800 GMT, 26 JANUARY 1961

north to south to the immediate west of Renfrew and passed through the station soon afterwards.

Local features.—The wind at 2000 feet estimated from the 1800 GMT chart was 200 degrees, 60 knots. The estimate from the 0001 GMT chart was 200 degrees, 45 knots. Table I shows the behaviour of the surface wind and outside air temperature.

Report of severe turbulence.—Aircraft landing during the evening experienced very considerable drift on the approach to runway 08 and all reported at least moderate turbulence between 500 and 1000 feet. Severe turbulence was reported by the pilot of the Viscount which landed at 2130 GMT. On his first approach, the pilot missed the runway altogether and gave his

TABLE I—SURFACE WIND AND TEMPERATURE AT RENFREW, 26–27 JANUARY 1961

Time	Wind			Temperature
GMT	degrees	knots		°C
1820	130	11		0.4
1850	120	15		0.4
1950	090	06		1.2
2050	150	07		1.7
2120	150	08		2.2
2150	160	06		2.3
2350	090	02		3.8
0020	190	19	Gust 30	3.4
0050	180	20	Gust 32	3.9

estimate of the wind at 1000 feet as 190 degrees, 45 knots, with degree of turbulence moderate. On his second approach, he "had great difficulty in controlling the aircraft in the circuit" and encountered "severe turbulence between 600 feet and 800 feet" at which heights the wind was estimated as 190 degrees, 40–45 knots. Below 500 feet, conditions became smooth, with little or no drift, and the aircraft landed on runway 26.

"With gradients of 160 degrees–210 degrees, winds above 40 knots will give normal surface directions and speeds." (Extract from "Notes on local weather at Renfrew"). In this case, however, the day started with surface temperatures at, or a little below, freezing-point, there was no sunshine and the maximum temperature between 0900 GMT and 2100 GMT was only 1.7°C. Thus, during the evening, a "cold pool" was established in the basin in which Renfrew lies; this

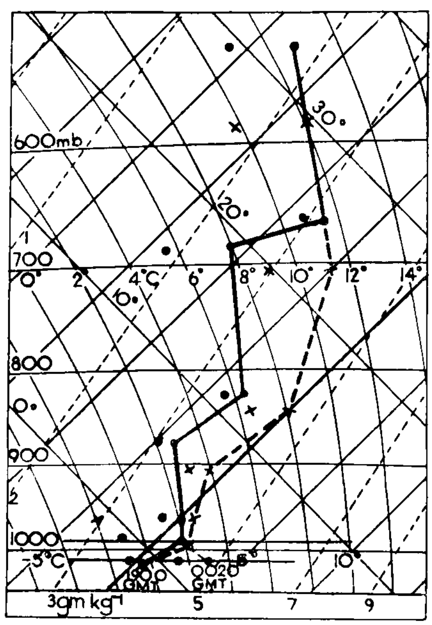

FIGURE 2—TEMPERATURE DISTRIBUTION WITH HEIGHT ON EVENING OF 26 JANUARY 1961

·————· Aldergrove, 1200 GMT, 26 January 1961

x – – – x Shanwell, 0001 GMT, 27 January 1961

Renfrew 1900 GMT surface temperatures have been substituted for the Aldergrove and Shanwell values, and the Renfrew 0020 GMT temperatures added.

pool remained undisturbed by the very strong gradient aloft. By substituting Renfrew 1900 GMT temperatures on an otherwise unaltered (and representative) Aldergrove 1200 GMT tephigram, this undisturbed region shows up as a slight inversion between the surface and 700 feet. The Shanwell sounding for midnight, unaltered except for the substitution of the same Renfrew surface temperatures, is of a pattern similar to that of the midday Aldergrove sounding, the layer between the surface and 700 feet being almost identical (see Figure 2).

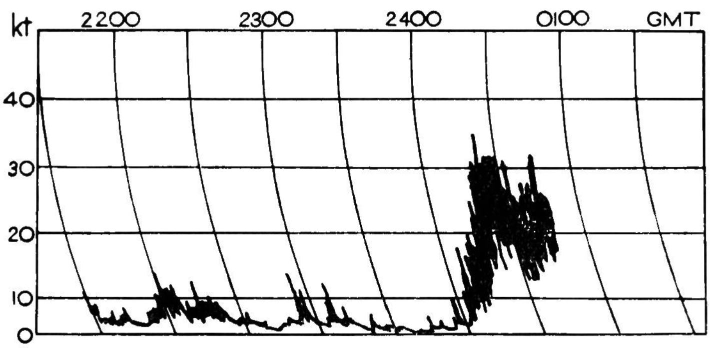

FIGURE 3—SURFACE WIND SPEED AT RENFREW, 26–27 JANUARY 1961

As warmer air progressed steadily eastwards, this inversion gradually broke down, finally allowing the strong winds to penetrate to the surface at 0020 GMT (see Figure 3). The 0020 GMT temperature is also shown in Figure 2.

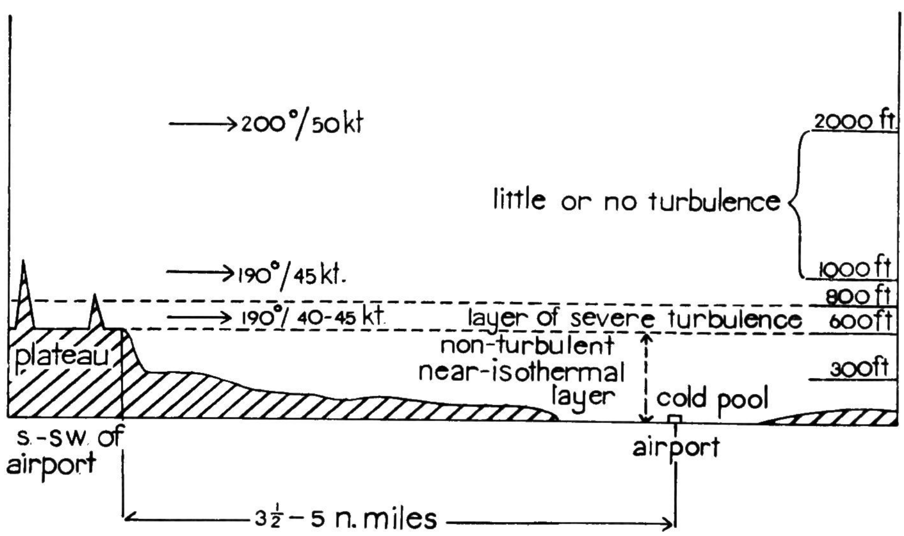

FIGURE 4—CROSS-SECTION SHOWING WIND DISTRIBUTION WITH HEIGHT AT RENFREW AT 2130 GMT, 26 JANUARY 1961

Figure 4 represents a cross-section of the air above the airport and surrounding district as the Viscount was preparing to land. The severe turbulence experienced was most likely due mainly to the abrupt wind shear immediately above the inversion, but frictional turbulence caused by the hills to the south and south-west of the airport may have been a contributory factor.

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ATMOSPHERIC DIFFUSION

By C. H. B. PRIESTLEY

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Given an amount of foreign substance (gaseous or particulate) released into the atmosphere in a known way, and given the condition of the atmosphere, how will that material become dispersed as a function of time? This is a subject of which a reasonably full and separate account now appears for the first time.* Those at all familiar with the complexity of the atmosphere will not expect any facet of the problem to be easy, and so it proves: the book is not for the lay reader. But it is a fascinating subject for the professional man, be he physicist, mathematician, or engineer; and there will be many outside meteorology who will require this book as a background for its applications in pollution and public health, in military and defence questions, and in the agricultural and botanical fields. Here is a book for which there is a real need, among so many for which the author's ego or pocket appears the main excuse.

A long review must both comment on the subject and describe the book, but Dr. Pasquill's arrangement is so logical that the two are best done together. To one who was introduced to atmospheric diffusion in 1939, towards the end of an era described by Dr. Pasquill as "the early days", the subject appeared already highly elaborated. But it has since acquired a radically new look, in the conception and development of which Dr. Pasquill and his colleagues at Porton have played leading roles. This has been an achievement which belongs by and large to the 1950 decade, so that the appearance of a connected narrative is both quick and timely.

In a short opening chapter we are introduced to the formal statistics which are used in general analysis of turbulence structure, most notably the autocorrelation and spectrum functions. We learn to look at these through a *window*, the slowest fluctuations being removed by sampling over a limited period τ, and the fastest by averaging over a period s. When the mean square value σ^2 of one of the components of turbulence is obtained in this way, it can be seen that $\sigma^2_{\tau,s}$ should be able to serve as an index of the dispersion of a cloud of material diffusing for a time s, from a source which emits for a time τ. That σ^2 goes on increasing, more or less indefinitely, with τ is the feature which distinguishes atmospheric from other forms of turbulence, and makes it the harder. The direct measurement of these window-framed functions, and the basing of diffusion estimates upon them, may be said to epitomize the new approach.

The next chapter summarizes the results of measurement of the spectrum and correlation functions in the atmosphere. Pause might be taken here to view the ultimate operational objective. Presumably at special places such as Windscale and Porton the need for window-spectrum information will be catered for by continuous monitoring and *ad hoc* analysis. But for general purposes one envisages the need for a climatology of these statistics, matching the comprehensive studies of vertical gradients, etc., in the early days by Johnson, Best, Flower and others. The task has been well started by Panofsky and his colleagues in the

**Atmospheric Diffusion*, by F. Pasquill. 9¼ x 6 in., pp. xii + 297, *illus.*, D. Van Nostrand Company Ltd., 358 Kensington High Street, London, W.14, 1961. Price: 60*s*.

United States of America, but much more remains to be done, both in the detail of the effects of thermal stratification, and in the atmosphere above the boundary layer. It is a formidable prospect in view of the doubly-infinite character of each element $\sigma^2_{\tau,s}$ and one can only hope that theory will come to the rescue and shorten the task by providing functional forms. Some attempts to this end are described, but few would regard them as definitive: even of the high-frequency properties, which appear the most amenable, Dr. Pasquill is compelled to write "the only simple rule yet to emerge even tentatively is that near the ground in neutral or unstable conditions the isotropic limit occurs at a wavelength roughly the same as the height above ground."

Chapter 3 sets out the main theory: not a theory of turbulence, but the formal development of the links between the chosen indices of the flow field which we can measure, and the dispersion which this flow field effects. Here we first veer away from purely statistical concepts and hark back to the eddy diffusivity K, to the profile of wind with height, to the differential equations akin to those of classical diffusion theory and their explicit solutions for the concentration in terms of x, y, z and t. This was a main theme of the early days, with Sutton as the outstanding exponent. Its essential limitation, long recognized, is the inability to handle all the spectrum attributes of the problem, in that a cloud will respond to different ranges of eddies as its own size continuously grows. Caught up in the enthusiasm for the new, more direct type of measurement, it may be too easy to overlook the major part that the old formulation, in terms of K and mean gradients, has played; and, extended by some flash of inspiration, may yet play again in development and understanding. Here only, for instance, does one find the physical reasons why the crosswind and vertical distributions of concentration are commonly sought as exponential functions. Again the ability to incorporate K as any function of position, and of height in particular, gives an inherent advantage in dealing with nature's reality, and we may look forward to further exploitation of this merit. Let us remember for example that a neutral or slightly unstable layer capped by an inversion is one of the most significant situations for pollution; and let us remember that the statistical theories are basically theories of a homogeneous field of turbulence.

The larger part of the chapter deals with the formal development of the statistical approach, from the foundations laid by G. I. Taylor and L. F. Richardson to the present day. This is the hardest part of the book, though master it one must if traps are to be avoided. Here too are fundamental obstacles yet to be surmounted. Whereas the particle responds to the fluctuations in velocity which it experiences as it moves along, i.e. in a Lagrangian framework, these are generally impracticable to measure and the bulk of data relates to fluctuations at a fixed point (the Eulerian system). It is intuitively clear that the time scale of the former should be the larger, and in practice the assumption of similarity of spectral form on a fourfold time scale has led in many instances to conformity with dispersion observations. But there is no evidence or argument to suppose that this represents a general truth.

And this is only the beginning of the trouble. The problem of the spread of a cluster involves essentially more than time variations for one point or one particle; new covariance spectra are required which relate the fluctuations of two particles both simultaneously and at different times. To obtain the instantaneous rate of growth we must look through yet another window—here called a band-pass filter—to cut out eddies which are either too large or too small to have expansionist tendencies: and the limits of the band must be allowed to change with time to match the changing size of the cluster. While the principles have been clarified the blue sky of practice has become very dark, and the data requirement is by now multiply infinite. The aim of theory will be to reduce this mass to a manageable number of parameters. But whilst we may admire the way such reductions have been achieved in the special case of isotropic turbulence, we dare not copy too closely because this is a simplification to which by and large, as every meteorologist knows, the atmosphere refuses to conform.

Chapter 4 pieces together the information coming from experiments aimed at answering salient questions such as:—What is the precise form of crosswind and vertical distribution? What index in the power law (with distance or elapsed time as the argument) best represents the variation of concentration or of dispersion for clouds of the idealized types (point source, line source, etc.). This is very thoroughly done and will be appreciated for reference purposes, though those concerned mainly with the flow of ideas may find the chapter somewhat tedious and may be excused for skipping. But it is necessary glue for coating the framework which has gone before so that practical prediction techniques may now be stuck on, however thin the glue may appear to be when it comes to other than near-neutral conditions. A valuable recapitulation concludes this, the more fundamental, part of the book.

As a preliminary to the problem of estimation we are next introduced to the main modes of behaviour of chimney and other plumes (looping, coning, fanning, lofting, fumigation) and the conditions in which each occurs. Illustrations of typical Dines traces would have been helpful in driving the lessons home, and it is to be regretted that the classification has not yet been taken beyond the descriptive stage. There follows a wholly admirable synthesis of the formula-techniques for estimating diffusion from each basic type of source, culminating in a general system of the author's own devising. This appears by judicious compromise to steer its way round many of the theoretical difficulties previously commented on. Having been led carefully through the details of the operation, in which he will need to keep his wits about him, the reader emerges equipped to give the answers if he knows the weather. How accurate the results will be, however, is left to his own judgment and experience and he may feel that the experts, by giving theirs, could have been more helpful on this point.

In the final chapter Dr. Pasquill turns from the necessarily idealized framework to consider some of the added complexities which arise in real situations. The first is the rise of a buoyant plume. Here theories are in existence and are reasonably adaptable but are still waiting for some good data, covering a range of the important atmospheric variables and measurements thereof. The account is accordingly scrappy, in contrast with the valuable synthesis of data on the spreading of chimney plumes which follows a little later. The intervening treatment of deposition deals with the concurrent processes of spreading and settling, with boundary condition subject to the concept of *deposition velocity*, or rate of deposition per unit area divided by the concentration. It is not clear that this concept, interpreted as it has been here in terms of the concentration at some distance from the surface, achieves anything more than the substitution of one phenomenon by another equally abstruse. The deposition velocity for instance is not equal to the terminal velocity but is dependent on the turbulent flow (including the nature of the boundary), an idea which is described as relatively novel but which should at once have been obvious from the kinship of the problem to that of natural evaporation. Technical blemishes in this section include the use of the same symbol, D, for two quite different quantities and a printing error and omission of units on the same line of page 235.

Finally there are brief summaries of the work on such varied applications as the surveys of distributions from single stacks, including the climatic distribution, the effects of gross surface irregularities, the Leicester survey, smog, the Windscale accident, bomb explosions, and the dispersal of spores and sprays and other minutiae of agricultural interest, including locusts. This, one might suspect, is the chapter which will most quickly call for amplification, if not for revision. Indeed there may well be an early need for a practical manual, dealing with the *ad hoc* problems in much greater detail than is possible here, and so forming a companion volume to the more basic considerations with which Dr. Pasquill has been mainly occupied. If we have to erect a stack in a given locality, where exactly shall we put it?

No reference whatever is made to the problems of evaporation from the Earth's surface which, energy consideration apart, are dominated by atmospheric diffusion and are probably the most important geophysical example thereof. In this process, unlike the applications treated by Dr. Pasquill, the properties of the source cannot be specified *a priori* since they are diffusion-dependent. As hinted earlier, this and other boundary complexities (the sometime presence of a laminar sublayer and the control of transpiration by stomata) intrude in much the same way as in the problem of deposition, and more contact between these fields might be stimulating to both.

And now what of the book, as distinct from its chapters? In assembling so much material for the first time Dr. Pasquill has had no easy task, and he has earned our admiration and our thanks for his clear connected account of the subject. This may be expected to take its place for some years as the standard work of reference. The style is mainly narrative and the author is punctiliously fair, at times even charitable, to the many who have worked along the way. At some points indeed, to obtain more critical and sharper focus, the general reader might have preferred a single composite re-analysis or re-synthesis of all the material by Dr. Pasquill himself. Where this has been done, as in Chapter 5, it is conspicuously successful. My only criticism of the arrangement is that the treatments of the two extraneous processes, fall-out and buoyant motion, are both unnecessarily split into two different parts of the book. The printing, diagrams, index and references are uniformly impressive.

And what in summary of the subject, as distinct from the book? Even in the early days it was recognized not only that knowledge of eddy intensity, or gustiness, was inherent for the prediction of diffusion but also that the relevant gustiness must be a function both of sampling time and, for a given cloud, of elapsed time. Nevertheless the hope was there that the main problem of that era, diffusion in the boundary layer, could be made determinate in terms of the bulk parameters alone (height, vertical wind and temperature gradient, roughness length). This is a justifiable scientific aspiration, and there still appears no reason why it should be abandoned. The march of events has, however, caused it to be set aside, because the interest in diffusion at higher levels has focussed the demand for a new look, while developments in computing and recording techniques have been timely in making this practicable. There have followed considerable gains in precision and, with the practical logic of bringing the predictor index closer to the predictee variable, presumably also a gain in accuracy when the data are available. This represents primarily an engineering rather than a fundamental advance. The clothes are now designed so that one can work most effectively with the tools of the day, but the body inside the clothes has changed but little. To mix the metaphor, in real understanding we still have to bridge the Lagrange–Euler ditch, which is deep though narrow, and the much wider river of non-isotropic intensity spectra and the influence of thermal stratification, etc. upon them. In the geophysical applications of atmospheric turbulence, where the concern is still with mean gradients and with flux rather than intensity spectra, rather more fundamental progress is evident. That this is because the problems are somewhat easier may provide the explanation but little solace, for there is danger that without the understanding the burst of progress which has followed the new approach to diffusion may now rather soon lose its momentum.

Moreover, conscious of the atmosphere's infinite variety, we must always be on guard lest any approach enjoying current success becomes too stereotyped. The global distribution of radioactivity from H-bombs, for example, depends on small but significant mean meridional motions, leakages through the break between troposphere and stratosphere, and other composite but not, in the usual sense, turbulent motions. And it is not only when one extends the *scale* that one meets a *type* of dispersing motion to which the normal formulations can never apply. We are reminded of some of these, though not all, at the opening to Chapter 5, where it is pointed out that the methods as they stand cannot be expected to give reliable estimates (*a*) in calm conditions, (*b*) when there are local disturbances, for example in the immediate vicinity of obstacles and (*c*) when the airflow is channelled or contains circulations or drainage set up by heating or hilly terrain. And how long is a piece of string? There are some questions, which were asked at Porton in the "early days" and are apparently still being asked, whose very nature permits no quantitative answer. Possibly these were in Dr. Pasquill's mind when he chose the enigmatic illustration on the dust-jacket of his book.

NOTES AND NEWS

High-level layer of dust

The photograph facing page 129 has been received from Dr. Bowen of the Commonwealth Scientific and Industrial Research Organization (CSIRO), Australia. Dr. Bowen's theories in relation to dust of meteoric origin are well known. His notes on the photograph are as follows:

This photograph was taken at 9.30 a.m. local time on 31 October 1961 by Major R. Anderson of the U.S.A.F. flying at an altitude of 66,000 feet at 46°S, 147°E, facing SE, and shows a marked dark band near the horizon. This band could be seen to the SE, S and SW, and persisted at least as far as 60°S. It was not visible from a height of 50,000 feet and optical considerations indicate that the layer must have been at a height close to that of the aircraft. On this occasion there was broken altostratus cloud to 20,000 feet and cirrus at 34,000 feet just below the tropopause.

This observation formed part of a programme of sampling of high-altitude aerosols on behalf of the Radiophysics Laboratory, CSIRO, Sydney. No such layers were observed visually on seven subsequent flights to similar heights.

LETTER TO THE EDITOR

Photographs of a radiosonde balloon

The accompanying photographs [between pp. 128–129] show the shape assumed by a radiosonde balloon before and after release. They show that the shape usually ascribed to balloons is no longer assumed when the balloon is in motion and at a sufficiently low altitude for it still to be very flabby. Ordinary pilot balloons are well stretched and are therefore not easily distorted. These photographs were taken at the Col des Aravis between Annecy and Mont Blanc in July 1960 where a team of French meteorologists, led by D. G. Barbé, was studying the small details of the air flow over the Alps by releasing balloons every hour for several days.

A patch of stratus can be seen in the ravine on the hill side.

Department of Mathematics, Imperial College, London, S.W.7. R. S. SCORER

REVIEW

Elements of dynamic meteorology, by A. H. Gordon, M.Sc. 8¾ in. x 6¼ in., pp. xxi + 217, *illus.*, The English Universities Press Ltd., 102 Newgate Street, London, E.C.1., 1962. Price: 25*s*.

In this book the author has attempted the very difficult task of providing a successor to Sir David Brunt's "Physical and dynamical meteorology". It has some good features but fails seriously.

Its first defect is the large amount of space devoted to utter trivialities. It may be possible to justify deriving the Clausius–Clapeyron equations, which is done with considerable care, but it is not reasonable, for example, to write the (two) equations for the geostrophic wind a second time merely with a term transposed to the other side. This is typical and the book abounds with examples.

Its major defect is in the poverty and occasional inaccuracy of its explanations of the physical significance of the treatment. One instance is that though the hydrostatic equation is derived (*sic*) and elaborated considerably, the essential consequence of its assumption, namely the removal of sound waves from the equations, is not mentioned. Another which may be quoted is in the discussion of the isallobaric wind. The isallobaric term becomes important when $\partial\mathbf{V}/\partial t$ is the major term in the expansion of the acceleration

$$\frac{\mathrm{d}\mathbf{V}}{\mathrm{d}t} = \frac{\partial\mathbf{V}}{\partial t} + \mathbf{V}.\nabla\mathbf{V}.$$

This does not mean "... that the (pressure) pattern is not moving in space ...". It means either that **V** is small or the gradient of **V** is small, or both. Such is the nature of the atmosphere that this usually means that **V** must be small. Judging by remarks later in the particular section of the book, this seems to be in accordance with Mr. Gordon's experience. These are examples picked at random from abundant material.

The usefulness of the book to many readers will be that quite a lot of the basic algebra of meteorology is duly derived. Mostly it is accurate, though the equation for the variation of wind with height looked deficient to the reviewer. But *please* can we recognize that the modern student is weaned on to vectors almost immediately after leaving school. If we use them they save a great deal of time and paper. In 1962 they should not be relegated to an appendix.

A. W. BREWER

OBITUARY

Mr. George Munn Gray Lightbody.—It is with deep regret that we learn of the death on 9 February 1962 of Mr. G. M. G. Lightbody, Senior Experimental Officer. George Lightbody joined the Office in January 1939 in the grade of Assistant III and was first appointed to Abbotsinch in the vicinity of his native Glasgow. On the outbreak of war he was transferred to Hornchurch, where he served during some of the worst days of the "blitz". On the brighter side, it was here also that he met his future wife.

After a period of training in forecasting duties at Gloucester he was promoted Acting Assistant II in 1942. Several postings to stations in East Anglia followed, and no doubt it was during this period that George acquired the facility of limpid clarity in briefing which was so characteristic of him. Further moves to Dunsfold and Netheravon followed, and eventually he was posted to India in December 1944. His promotion to Acting Assistant I was promulgated while he was still en route to the east.

Returning to England in 1946, he was established in the grade of Experimental Officer in 1947 and served for several years at Northolt before going abroad again to Cyprus in 1951. A serious illness in Cyprus marred, but did not destroy, his enjoyment of the then peaceful and delightful island, nor did it deter him from plunging whole-heartedly into the activities of the *ad hoc* committee formed there to draw up the first foreign service allowance price returns, to which committee he was appointed as Air Ministry representative.

Home again in 1954, he went to Leeming, where he remained until November 1960, when he was promoted Senior Experimental Officer and posted to Uxbridge, where he remained until his untimely death.

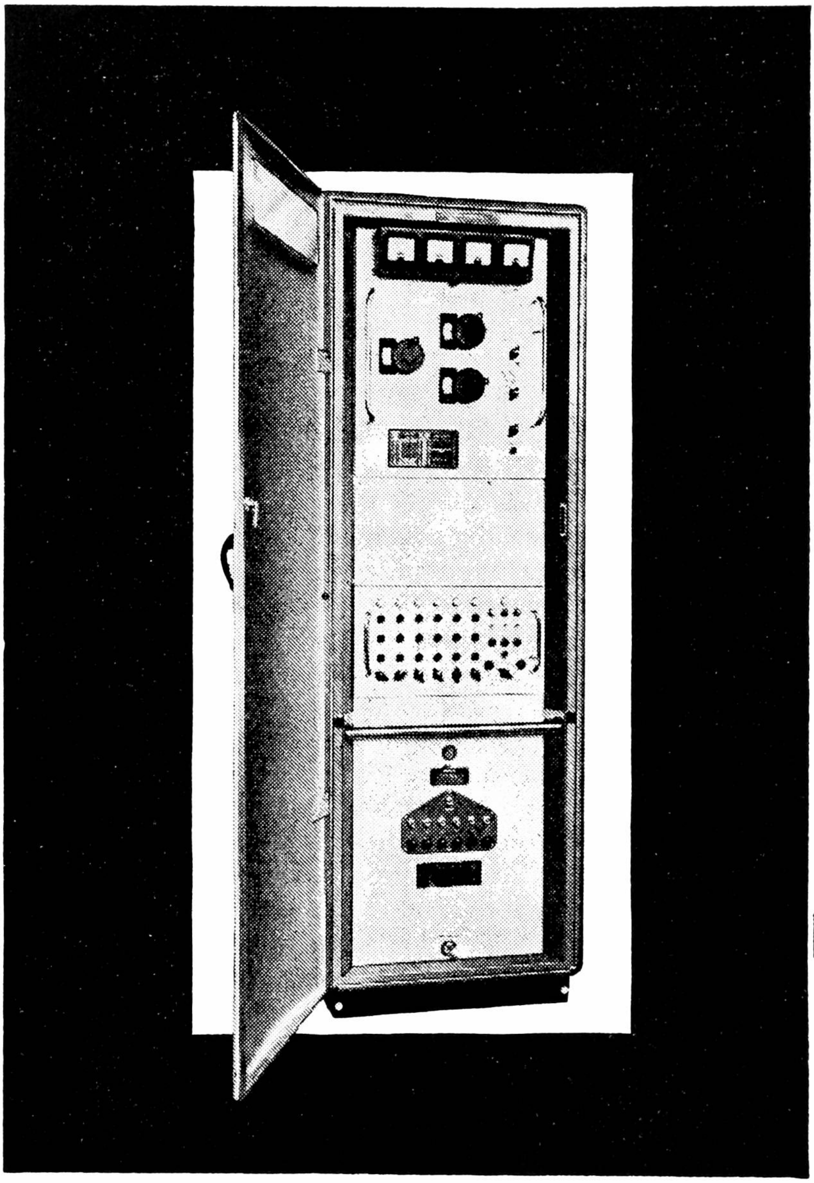
George was always a genial and gregarious person and a good friend. His many longstanding friendships amongst his colleagues and also in the Royal Air Force bear witness to his personal sincerity and charm, and no less to the esteem which his high professional standards inspired. His zest was infectious, and there will be few who knew him who will not most readily remember some of his good companionship or perhaps some characteristic act of understanding and generosity. All his colleagues and friends will wish to extend their heartfelt sympathy to his widow and two children.

L.J.A.

METEOROLOCICAL OFFICE NEWS

Report on the first season of the Meteorological Office, Bracknell, Table Tennis Club

The Table Tennis Club was formed at the beginning of this season and entered a team in the Third Division and another in the Fourth Division of the local league. Much to the surprise of the original members both teams went through the season undefeated and emerged as winners of their respective leagues, collecting handsome trophies. Next season it is hoped to run three, or perhaps four, teams and we hope to maintain our run of success after promotion to higher divisions.



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