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THE USE OF WET-BULB POTENTIAL TEMPERATURE CHARTS

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SUMMARY

Charts showing the distribution of wet-bulb potential temperature (θ_w) on isobaric surfaces were found to be a valuable aid in analysis and forecasting. Charts of θ_w at 850 mb show low-level air-mass types more clearly than other charts at present in use and enable frontal analysis to be improved. There was found to be a relationship between the value of the 850 mb θ_w and the probability of precipitation falling as snow. High values of the 850 mb θ_w were observed to be associated with the development of summer thunderstorms. Charts of θ_w at 500 mb and 850 mb were combined to derive an index of potential stability. Charts showing the variation of this stability index over a wide area were found to be useful in defining areas where thunderstorms or significant falls of rain were likely.

INTRODUCTION

For many years the major aids for analysis and prognosis were the contour charts of the 500 mb surface and the thickness of the 1000/500 mb layer. When all the work was done by hand, much of the available time was spent in plotting, analysing and constructing forecast charts for these levels. Now that the majority of surface and upper-air reports are received, checked and stored by computers, which can then produce mechanically plotted and objectively analysed charts, the forecaster has the opportunity to examine the atmosphere in greater detail. There is a limit to the number of charts which can usefully be examined, but it is suggested that charts which summarize the distribution of moisture and stability in the troposphere can form a valuable addition to the present routine.

Wet-bulb potential temperatures* (θ_W) combine information on temperature and vapour content in a single figure which is conservative for adiabatic changes. A chart showing isopleths of θ_W on a single isobaric surface can be used to improve air-mass and frontal analysis. From the distribution of θ_W on two isobaric surfaces one may derive an index of potential stability which can be

* Because of the inevitable errors introduced by the methods of approximation and calculation employed, the theoretical distinction between 'wet-bulb potential temperature' and 'pseudo wet-bulb potential temperature' is of no importance in the present work and the former term will be used for convenience.

used over a wide area. When such a chart is used in conjunction with standard surface and upper-air contour charts, the regions where moderate rain or thunderstorms are likely may be defined with increased confidence.

SELECTION OF ISOBARIC SURFACES FOR θ_W CHARTS

The 850 and 500 mb surfaces were found to be the most useful standard levels for plotting the distribution of θ_W over a sector of the northern hemisphere extending from North America to central Russia. The 850 mb surface was generally high enough to be little influenced by diurnal variations but low enough to be representative of conditions near the base of the troposphere. The 500 mb surface was the highest that could be used throughout the year. Higher levels were above the tropopause for periods during the colder months.

In summer months and in low latitudes the 400 mb or even the 300 mb surface appeared useful for locating areas of deep instability, but these levels suffer from errors due to the decrease of accuracy in the reported values of temperature and humidity. These radiosonde errors are generally found to increase with height.

CONSTRUCTION OF θ_W CHARTS

The values of θ_W can be found graphically by using a tephigram but the process is far too slow for routine work. For operational purposes the extraction of data is best handled by the computer, but for experimental work, and when the necessary data are not accessible by machine, the values of θ_W should be read from a Table. Table I allows the user to find the 850 mb θ_W from the temperature and dew-point depression reported at that level in part A of routine radiosonde messages. In this Table the values have been rounded to the nearest whole degree. Values ending in ·5 have been rounded up. For most purposes this Table is adequate since the original observations rarely justify working to a greater degree of precision.

Over most of Europe, western Russia and North America the network of radiosonde stations is close enough for a preliminary sketch of θ_W isopleths to be drawn by direct interpolation aided by continuity. Over the oceans these methods need to be supplemented by extrapolation, guided by the use of satellite data. Preliminary 850 mb isopleths should be superimposed on the first drawing of the surface chart and adjustments made to both to obtain consistency.

Isopleths of θ_W at 500 mb show less direct relationship to the surface chart, and the general pattern bears more resemblance to the contours of the 300 mb surface. The differences are most noticeable over occlusions, which are often marked by a warm tongue in the θ_W isopleths, and near major vortices when the 300 mb low is not concentric with lower-level features. Isopleths showing the difference in θ_W between 850 and 500 mb can provide an index of potential instability (or stability). These lines can be drawn directly, using the values determined from each sounding, but better results may be obtained by gridding the two levels on a light-table. The technique is similar to that employed when contour charts are constructed with the aid of partial thicknesses.

The analyst needs to watch for two common errors in the data. The first is due to a coding mistake which causes the temperature to be read as negative when it should be positive, or vice versa. These coding errors usually produce such a gross distortion of the isopleths that the mistake becomes obvious. The second type of error occurs when the radiosonde drifts off calibration during the ascent. Values at 500 mb and higher are more sensitive to this kind of error than are the

850 mb values. Fortunately errors in θ_W are usually in the same sense as errors in the heights of the 300 or 500 mb surface. Anomalously high values of the 500 mb θ_W generally coincide with reported contour heights which are noticeably higher than surrounding values, and this feature may be used to correct or reject the report.

USE OF 850 mb θ_W ISOPLETHS

The primary use of these charts is for air-mass and frontal analysis. Frontal analysis is still a subjective process, as can be seen when charts issued by different meteorological centres are compared. The difficulty of satisfactory analysis is increased when the charts in use are on a small scale with only a limited number of observations plotted. The isopleths of the 850 mb θ_W can help the analyst to place fronts with a clearer knowledge of the change of air mass at the frontal boundaries.

Where surface fronts were well defined, the isopleths of the 850 mb θ_W and the contours of the 1000/500 mb thickness showed very similar patterns. In these cases the surface observations alone were usually sufficient to define the frontal position. Where the surface fronts were poorly defined by surface observations, the 850 mb θ_W isopleths appeared superior to the 1000/500 mb thickness lines in locating the position and alignment of the fronts. The θ_W isopleths enabled an analyst to trace the movements of air masses during periods when the front separating them was too inactive to be located by surface observations or cloud patterns. In some cases frontogenesis was apparent some 12 hours earlier on θ_W charts than on surface or 1000/500 mb thickness charts.

850 mb θ_W ISOPLETHS AND FRONTS

The following relationships between θ_W isopleths and surface fronts were observed:

(a) The isopleths lay almost parallel to warm and cold fronts for very long distances even when the fronts were aligned north to south.

(b) The gradient of θ_W was greatest in the cold air, and the position of the surface front often coincided with a marked decrease in this gradient. Figure 1 shows an example of this change of gradient to the south of a front which lay from the northern Caspian Sea to Scandinavia and beyond.

(c) The maximum values of θ_W often occurred close to the surface front on the warm side of the boundary. Figure 2 shows an example of this. The appearance of a very narrow, warm, moist tongue in such a position was observed on many occasions. It appeared to fit the model of a low-level jet stream described by Browning (1973). This jet stream consisted of a tongue of anomalously warm moist air where θ_W attained a maximum value in the horizontal.

(d) Where occlusion had occurred, the isopleths of θ_W lay across the line of the front. An example is shown in Figure 3. An old occlusion lay from Leningrad to north-west Germany and a newly formed occlusion had developed off the east coast of Greenland.

NON-FRONTAL PATTERNS OF θ_W

Isopleths of the 850 mb θ_W proved useful in showing when the air mass was not homogeneous. The air in warm sectors was generally more nearly homogeneous than the air behind cold fronts, but tongues of drier air were observed in a

TABLE I—TO DETERMINE 850 mb WET-BULB POTENTIAL TEMPERATURE, USING AIR TEMPERATURE AND DEW-POINT DEPRESSION

Air temp. (°C)	Dew-point depression (°C)																					
	00	01	02	03	04	05	06	07	08	09	10	11	12	13	14	15	16	17	18	19	20	25	30
30	35	34	34	33	32	31	31	30	29	29	28	28	27	27	26	26	25	25	24	24	23	22	20
29	34	33	33	32	31	31	30	29	29	28	27	27	26	26	25	25	24	24	23	23	23	21	20
28	33	32	32	31	30	30	29	28	28	27	27	26	26	25	25	24	24	23	23	22	22	20	19
27	32	32	31	30	29	29	28	28	27	26	26	25	25	24	24	23	23	22	22	22	21	20	18
26	31	31	30	29	29	28	27	27	26	26	25	24	24	23	23	23	22	22	21	21	21	19	18
25	30	30	29	28	28	27	26	26	25	25	24	24	23	23	22	22	21	21	21	20	20	18	17
24	29	29	28	27	27	26	26	25	24	24	23	23	22	22	22	21	21	20	20	20	19	18	17
23	29	28	27	27	26	25	25	24	24	23	23	22	22	21	21	20	20	20	19	19	19	17	16
22	28	27	26	26	25	24	24	23	23	22	22	21	21	20	20	20	19	19	19	18	18	17	16
21	27	26	25	25	24	24	23	23	22	21	21	21	20	20	19	19	19	18	18	18	17	16	15
20	26	25	24	24	23	23	22	22	21	21	20	20	19	19	19	18	18	17	17	17	16	15	14
19	25	24	24	23	22	22	21	21	20	20	19	19	19	18	18	17	17	17	17	16	16	15	14
18	24	23	23	22	22	21	21	20	20	19	19	18	18	17	17	17	16	16	16	16	15	14	13
17	23	22	22	21	21	20	20	19	19	18	18	17	17	17	16	16	16	15	15	15	15	14	13
16	22	22	21	20	20	19	19	18	18	18	17	17	16	16	16	15	15	15	14	14	14	13	12
15	21	21	20	20	19	19	18	18	17	17	16	16	16	15	15	15	14	14	14	14	13	12	11
14	20	20	19	19	18	17	17	17	16	16	16	15	15	15	14	14	14	13	13	13	13	12	11
13	19	19	18	18	17	17	16	16	16	15	15	14	14	14	14	13	13	13	12	12	12	11	11
12	19	18	17	17	17	16	16	15	15	14	14	14	13	13	13	13	12	12	12	12	11	11	10
11	18	17	17	16	16	15	15	14	14	14	13	13	13	12	12	12	12	11	11	11	11	10	9
10	17	16	16	15	15	14	14	14	13	13	13	12	12	12	11	11	11	11	10	10	10	9	9
09	16	15	15	14	14	14	13	13	12	12	12	12	11	11	11	10	10	10	10	10	9	9	8
08	15	14	14	14	13	13	12	12	12	11	11	11	11	10	10	10	10	9	9	9	9	8	8
07	14	14	13	13	12	12	12	11	11	11	10	10	10	10	9	9	9	9	8	8	8	8	7
06	13	13	12	12	12	11	11	10	10	10	10	9	9	9	9	8	8	8	8	8	8	7	6
05	12	12	11	11	11	10	10	10	9	9	9	9	8	8	8	8	8	7	7	7	7	6	6
04	11	11	11	10	10	10	9	9	9	8	8	8	8	7	7	7	7	7	6	6	6	6	5
03	11	10	10	9	9	9	8	8	8	8	7	7	7	7	6	6	6	6	6	6	6	5	5
02	10	9	9	9	8	8	8	7	7	7	7	6	6	6	6	6	5	5	5	5	5	4	4
01	9	8	8	8	7	7	7	7	6	6	6	6	5	5	5	5	5	5	4	4	4	4	3
00	8	8	7	7	7	6	6	6	6	5	5	5	5	4	4	4	4	4	4	4	4	3	3
−01	7	7	6	6	6	5	5	5	5	5	4	4	4	4	4	3	3	3	3	3	3	2	2
−02	6	6	6	5	5	5	4	4	4	4	4	3	3	3	3	3	3	2	2	2	2	2	2
−03	5	5	5	4	4	4	4	3	3	3	3	3	2	2	2	2	2	2	2	2	1	1	1
−04	4	4	4	4	3	3	3	3	2	2	2	2	2	2	1	1	1	1	1	1	1	0	0

−05	3	3	3	3	2	2	2	2	2	1	1	1	1	1	1	1	0	0	0	0	0	0	−1
−06	3	2	2	2	2	1	1	1	1	1	0	0	0	0	0	0	0	0	−1	−1	−1	−1	−1
−07	2	2	1	1	1	1	0	0	0	0	0	−1	−1	−1	−1	−1	−1	−1	−1	−1	−1	−2	−2
−08	1	1	0	0	0	0	−1	−1	−1	−1	−1	−1	−1	−2	−2	−2	−2	−2	−2	−2	−2	−3	−3
−09	0	0	−1	−1	−1	−1	−1	−2	−2	−2	−2	−2	−2	−2	−2	−3	−3	−3	−3	−3	−3	−3	−3
−10	−1	−1	−1	−2	−2	−2	−2	−2	−2	−3	−3	−3	−3	−3	−3	−3	−3	−3	−3	−4	−4	−4	−4
−11	−2	−2	−2	−2	−3	−3	−3	−3	−3	−3	−4	−4	−4	−4	−4	−4	−4	−4	−4	−4	−4	−5	−5
−12	−3	−3	−3	−3	−4	−4	−4	−4	−4	−4	−4	−4	−5	−5	−5	−5	−5	−5	−5	−5	−5	−5	−5
−13	−4	−4	−4	−4	−4	−5	−5	−5	−5	−5	−5	−5	−5	−5	−6	−6	−6	−6	−6	−6	−6	−6	−6
−14	−5	−5	−5	−5	−5	−5	−6	−6	−6	−6	−6	−6	−6	−6	−6	−6	−6	−6	−7	−7	−7	−7	−7
−15	−5	−6	−6	−6	−6	−6	−6	−6	−7	−7	−7	−7	−7	−7	−7	−7	−7	−7	−7	−7	−7	−8	−8
−16	−6	−7	−7	−7	−7	−7	−7	−7	−7	−8	−8	−8	−8	−8	−8	−8	−8	−8	−8	−8	−8	−8	−8
−17	−7	−7	−8	−8	−8	−8	−8	−8	−8	−8	−8	−9	−9	−9	−9	−9	−9	−9	−9	−9	−9	−9	−9
−18	−8	−8	−8	−9	−9	−9	−9	−9	−9	−9	−9	−9	−9	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10
−19	−9	−9	−9	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10	−10	−11	−11	−11	−11	−11	−11
−20	−10	−10	−10	−10	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−11	−12	−12
−21	−11	−11	−11	−11	−11	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12	−12
−22	−12	−12	−12	−12	−12	−12	−12	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13	−13
−23	−13	−13	−13	−13	−13	−13	−13	−13	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14
−24	−14	−14	−14	−14	−14	−14	−14	−14	−14	−14	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15
−25	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−15	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16
−26	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−16	−17	−17	−17	−17	−17	−17
−27	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−17	−18	−18
−28	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18	−18
−29	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19	−19
−30	−19	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20	−20

Note: Values have been rounded to the nearest whole number. Values ending in ·5 have been rounded up.

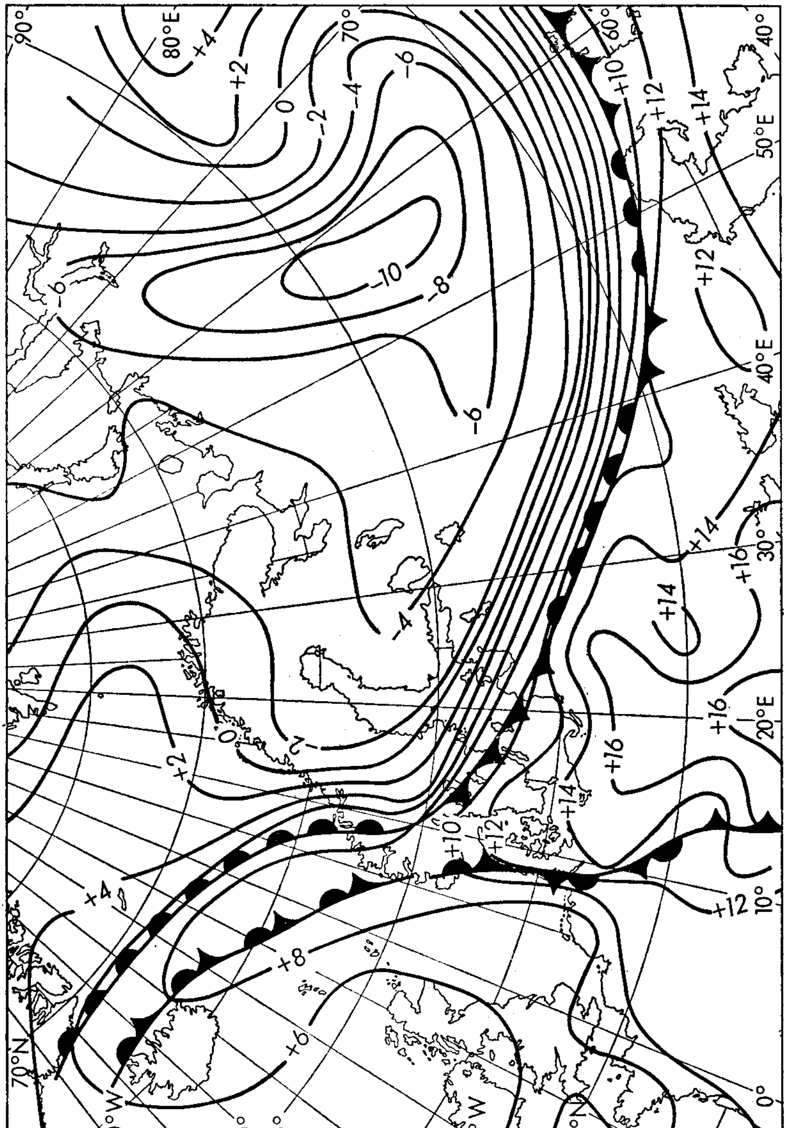

FIGURE 1—850 mb θ_W AT 1200 GMT ON 13 OCTOBER 1976, WITH SURFACE FRONTS, SHOWING STRONG GRADIENT OF θ_W IN THE COLD AIR NORTH OF THE FRONT EXTENDING FROM THE NORTHERN CASPIAN SEA TO SCANDINAVIA

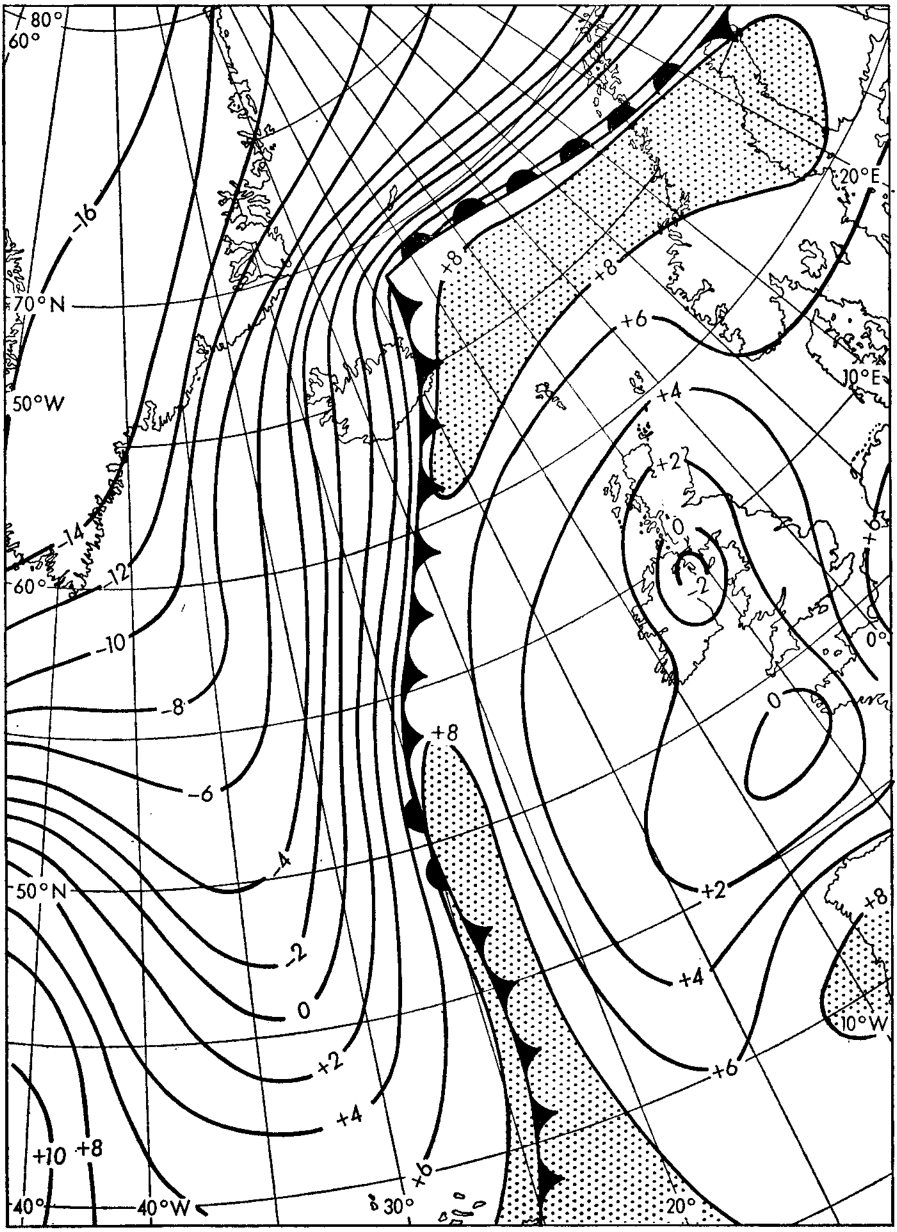

FIGURE 2—850 mb θ_w AT 0000 GMT ON 17 FEBRUARY 1976, SHOWING TONGUE OF WARM MOIST AIR JUST AHEAD OF THE SURFACE FRONT

Areas with values of θ_w of 8°C or more are stippled.

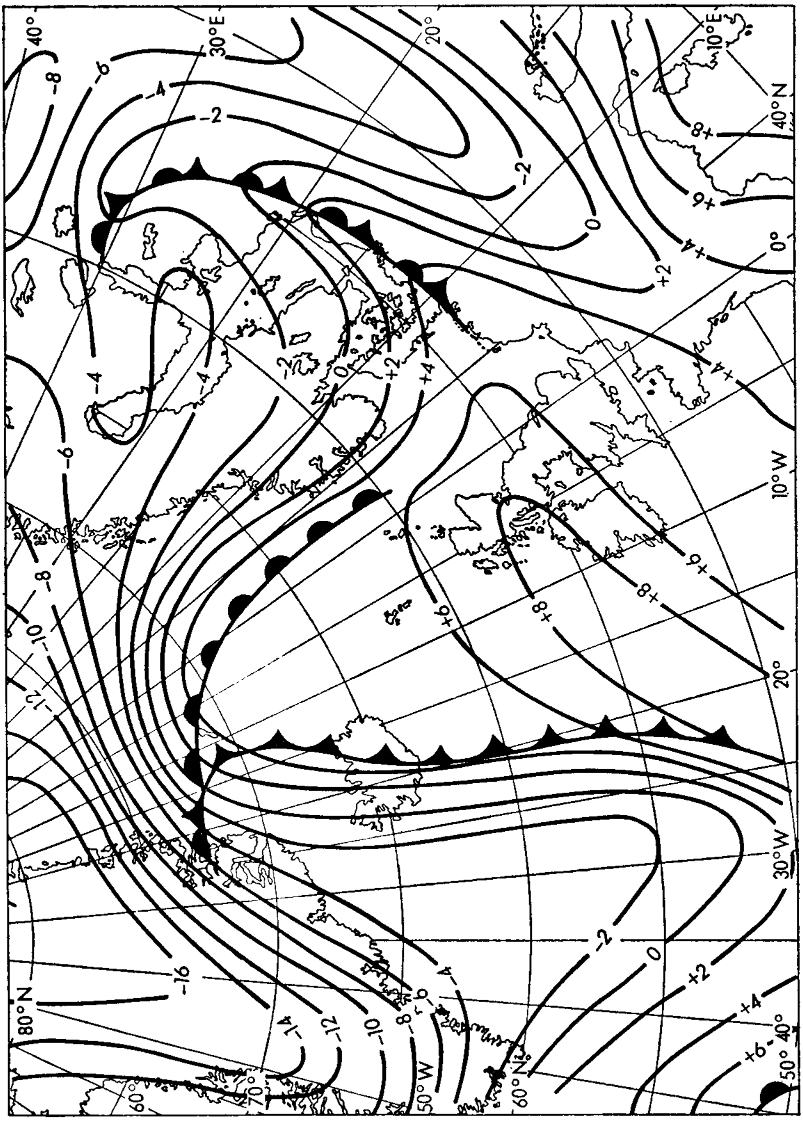

FIGURE 3—850 mb θ_W AT 1200 GMT ON 19 DECEMBER 1975, SHOWING ISOPLETHS CROSSING OCCLUDED FRONT BUT LYING PARALLEL TO WARM AND COLD FRONTS

number of warm sectors. Such dry tongues were usually observed when the warm sector lay on the perimeter of a surface anticyclone. These dry tongues were usually associated with a marked reduction in the amount of low cloud.

NON-ADIABATIC CHANGES OF θ_w

Although θ_w is insensitive to adiabatic changes, it is very sensitive to changes in the vapour content of the air. Rapid changes were observed when cold dry air from the Arctic was advected across a relatively warm sea. Vigorous convective currents rising from the sea surface carried heat and moisture above the 850 mb surface, and θ_w was observed to increase by more than 20°C in air which moved from Canada to western Europe in winter.

Decreases in the value of θ_w were observed to occur more slowly than increases. The most common cause for a fall in the value of θ_w was prolonged precipitation falling from a warm air mass moving over a cold surface. When unusually strong winds brought air from the western North Atlantic to Ellesmere Island, north-west of Greenland, the 850 mb θ_w fell from about 15° to 5°C during its passage from 45°N to 80°N. This represents about half the change which was normally observed when the flow was in the opposite direction in winter.

The value of θ_w as a tracer of air masses is inevitably reduced by these non-adiabatic changes, but such changes enable the analyst to observe the modifications which have taken place in the lower levels of the troposphere.

850 mb θ_w AND SNOWFALL

Lowndes *et alii* (1974) found that the wet-bulb freezing level was the best single predictor for use in deciding whether precipitation would fall as snow. The isopleths of the 850 mb θ_w can also be used for this purpose. The indications are less accurate than the wet-bulb freezing level but are easier to use over a broad area. Table II shows the results from a survey of over 1000 observations of precipitation which occurred at, or very close to, a radiosonde station at the time of the soundings. It was observed that the chief reason for the wide range of values was variations in the lapse rate of θ_w. When the air had passed over a relatively warm sea, precipitation at stations near the coast was of rain rather than snow even when the value of the 850 mb θ_w was close to 0°C. In contrast, areas which were already snow covered, such as eastern Europe and Russia, experienced snow rather than rain even though the value of the 850 mb θ_w was high, if the warm air had had a long passage over a frozen surface.

TABLE II—850 mb θ_w AND THE PROBABILITY OF PRECIPITATION FALLING AS SNOW

	Wet-bulb potential temperature (θ_w) in degrees Celsius										
	9	8	7	6	5	4	3	2	1	0	—1
Probability of snow	0	0·01	0·10	0·14	0·26	0·39	0·57	0·88	0·90	0·96	1·00

850 mb θ_w AND THUNDERSTORMS

Endlich and Mancuso (1968) reported that low-level temperature and moisture fluxes are more directly related to severe storms than are the lapse rate and parcel instability. David (1976) observed that one of a number of parameters associated with severe storms which produced tornadoes in North America was a mean 850 mb temperature of 15°C with a dew-point of 9°C. These figures give a value for θ_w of 18°C.

Although tornadoes are rare in the British Isles and north-west Europe, the incidence of severe summer thunderstorms appeared to be associated with high values of the 850 mb θ_W. Many summer thunderstorms broke out when southerly winds at low levels had advected air with 850 mb θ_W of 16°C or more over Europe, particularly when this warm moist air lay to the east of an advancing upper trough. A number of storms developed near to, or just west of, the centre line of a warm ridge in the 1000/500 mb thickness pattern, a region which might not be supposed to be particularly unstable.

Figures 4, 5 and 6 show examples of thunderstorms which were associated with high values of the 850 mb θ_W. The positions of independently analysed surface fronts have been marked. Thunderstorms reported in synoptic messages or located by the SFLOC network are marked by lightning symbols.

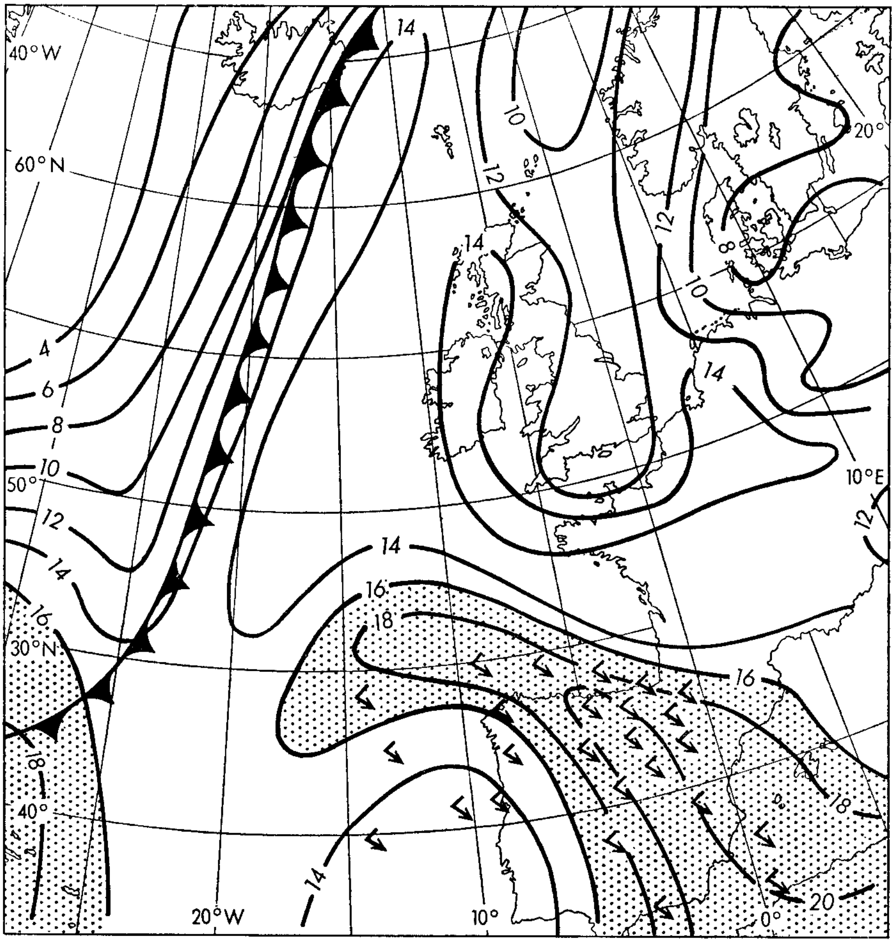

FIGURE 4—850 mb θ_W AT 1200 GMT ON 21 AUGUST 1976, SHOWING POSITIONS OF LIGHTNING FLASHES OBSERVED OR REPORTED BY SFLOC MESSAGES BETWEEN 1200 AND 1730 GMT

Areas with values of θ_W of 16°C or more are stippled.

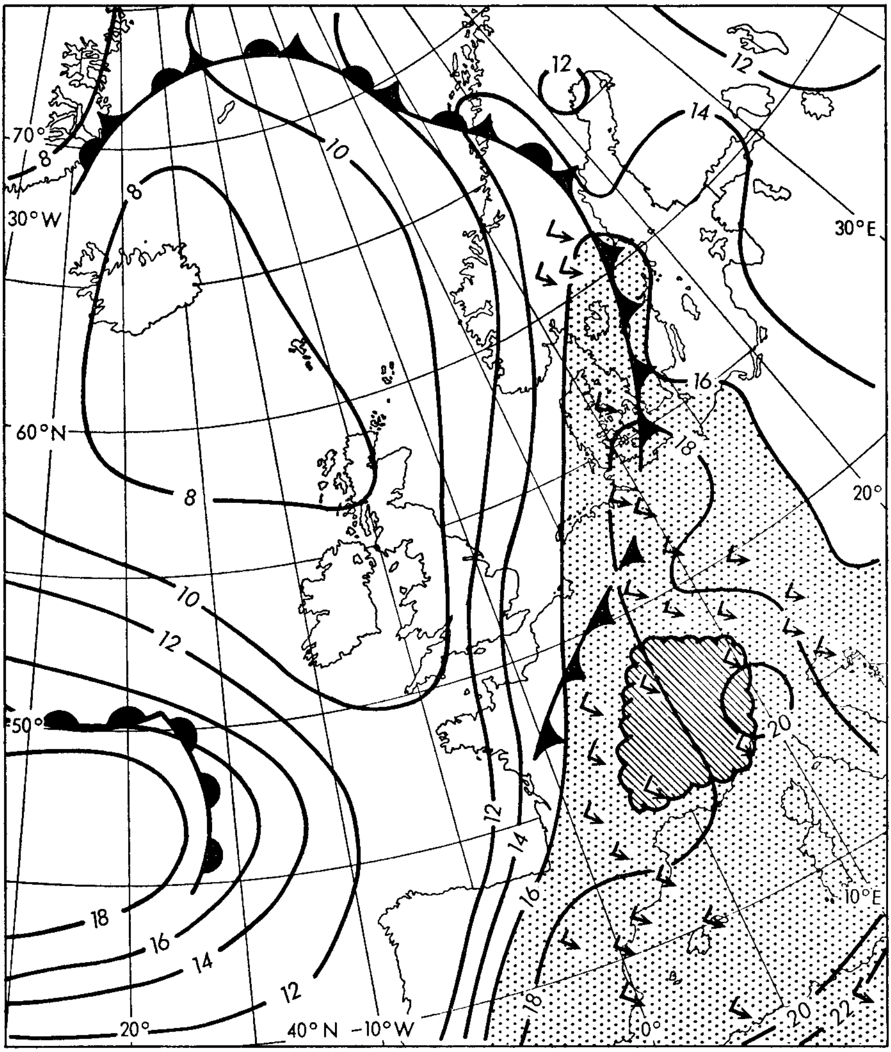

FIGURE 5—850 mb θ_w AT 1200 GMT ON 17 JULY 1976, SHOWING POSITIONS OF LIGHTNING FLASHES OBSERVED OR REPORTED BY SFLOC MESSAGES BETWEEN 1200 AND 1730 GMT

An area of particularly numerous reports over Switzerland and south-east France is shaded. Areas with values of θ_w of 16°C or more are stippled.

A POTENTIAL STABILITY INDEX

The 850 mb θ_w alone failed to show up occasions when thunderstorms occurred in a relatively cool air mass. The use of a potential stability index was found to help in locating areas where the thunder risk was high. This index was obtained by subtracting the value of θ_w at 850 mb from the value at 500 mb. When this figure was negative the air between the two levels was potentially unstable.

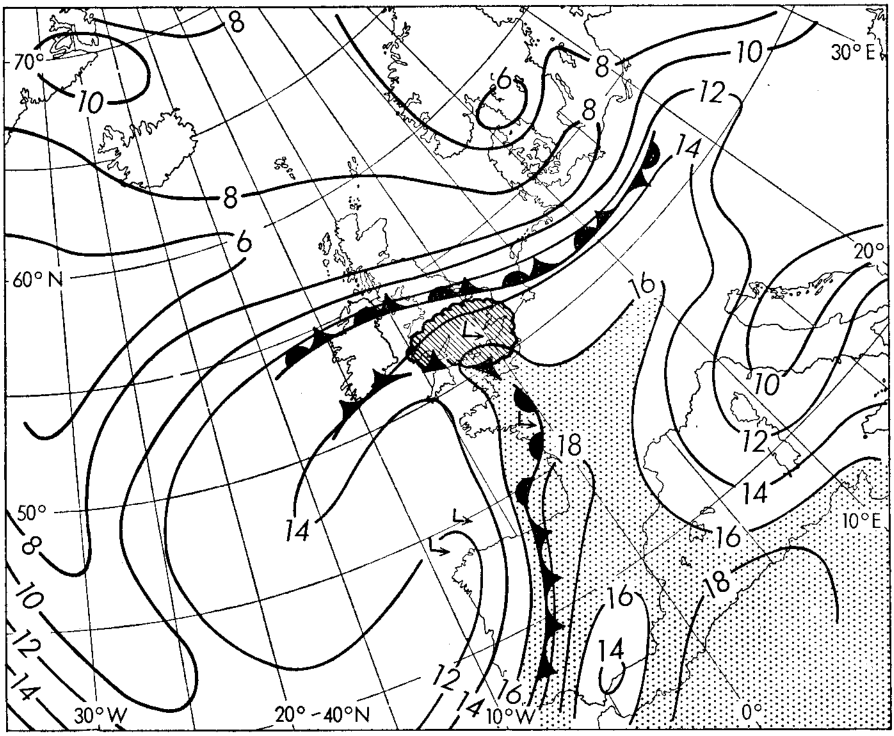

FIGURE 6—850 mb θ_W AT 0000 GMT ON 25 SEPTEMBER 1976, SHOWING AN AREA OF VERY NUMEROUS THUNDERSTORMS WHICH OCCURRED BETWEEN 0000 AND 0530 GMT OVER SOUTHERN ENGLAND

Areas with values of θ_W of 16°C or more are stippled.

Similar indices have been used by Rackliff (1962) and Jefferson (1963) but these differ in using the actual temperature at 500 mb instead of θ_W. They also used the 900 mb θ_W instead of the 850 mb value, but this change was regarded as unimportant, provided that there was no inversion between the two levels.

In order to determine the values of potential stability on thunderstorm days, 544 soundings made between 1973 and 1976 were examined. The results are shown in graphical form in Figure 7. The line marked 'Limit' shows the most unstable cases, and the line marked '5%' shows the values above which thunderstorms were rarely observed.

The slope of these lines shows that the critical value of the potential stability index varied with the 850 mb θ_W. During the summer months, when the 850 mb θ_W was relatively high, thunderstorms were observed only if the air was potentially unstable. In cooler periods, when the 850 mb θ_W was low, thunderstorms were observed when the index showed a few degrees of stability.

A similar result was found when the Jefferson index was calculated, using the formula $T_J = 1{\cdot}6\ (\theta_{W850} - T_{500} - 11)$. However, the Jefferson index showed a greater variability over its range. The standard deviation for the Jefferson index was 2·5 compared to 1·38 for the potential stability index.

It seems probable that the change in the critical values of both stability indices over the temperature range was due to different sizes of cumulonimbus. In winter, when the freezing level is low, thunderstorms can occur in cumulonimbus clouds whose summits do not reach the 500 mb level, or only exceed it by a small margin. In summer, when the freezing level is high, the tops of thundery cumulonimbus often extend far above the 500 mb level.

POTENTIAL STABILITY AND NON-THUNDERY RAIN

During the survey of the relationship between the potential stability index and thunderstorms it was noticed that the index was frequently low on occasions when rain fell but no thunder was reported. A second survey was made to include all occasions when precipitation, other than drizzle, was reported at, or very close to, a radiosonde station at the time of the sounding. Figure 8 shows the results from 800 soundings made in precipitation areas, superimposed on the graph of thundery soundings for comparison. As might be expected, the lines marked '50% ppn' and '5% ppn' show greater stability than the thunderstorm lines, but the difference between the lines decreases as the value of the 850 mb θ_w rises. When the 850 mb θ_w reaches 18°C, the two lines marked '50%' are less than 1°C apart. The graph of non-thundery occasions was not taken beyond this point because at and above 18°C almost all examples were associated with thunder at some stage.

INADEQUACY OF A SINGLE INDEX FOR THUNDERSTORM PREDICTION

It must be stressed that Figure 7 does not provide a thunderstorm index. The figures on which it is based do not show the occasions when the air was potentially unstable but no thunder occurred. It is difficult to establish with certainty that thunder did not occur, and no attempt was made to do so. The graph merely indicates a range of conditions which existed when thunderstorms did occur: from this one may infer that thunderstorms are unlikely outside the range, but not that thunder will occur if conditions lie within the range.

Saunders (1966 and 1967) tested various indices for thunderstorm prediction and found that none produced results as good as the subjective assessments made by forecasters who used indices as a guide rather than as a strict rule. It is clearly essential to use such objective guides in combination with surface and upper-air charts. Table III lists a number of features which appeared to be most often associated with thunderstorms. The Table suggests that surface heating was not by itself sufficient to account for many thunderstorms and that the low-level convergence of moisture provided by a trough was important.

TABLE III—PROBABILITY THAT A THUNDERSTORM WILL BE ASSOCIATED WITH PARTICULAR FEATURES ON SURFACE AND 300 mb CHARTS

Surface charts	*Prob.*	300 mb charts	*Prob.*
Near a trough or low	0·70	Near or ahead of a trough or low	0·76
With general cyclonic curvature	0·11	Behind a mobile ridge or high	0·08
Col or indefinite pattern	0·09	Behind a mobile trough or low	0·06
Straight isobars	0·04	Near a ridge or high	0·05
Anticyclonic curvature	0·05	Other features	0·05
Near a high or marked ridge	0·01		

Notes: (a) SFLOC reports were not included in this Table.
(b) The area examined extended from the British Isles to Poland, and from northern Italy to southern Scandinavia.
(c) Fronts are included under troughs, provided that the two coincided.

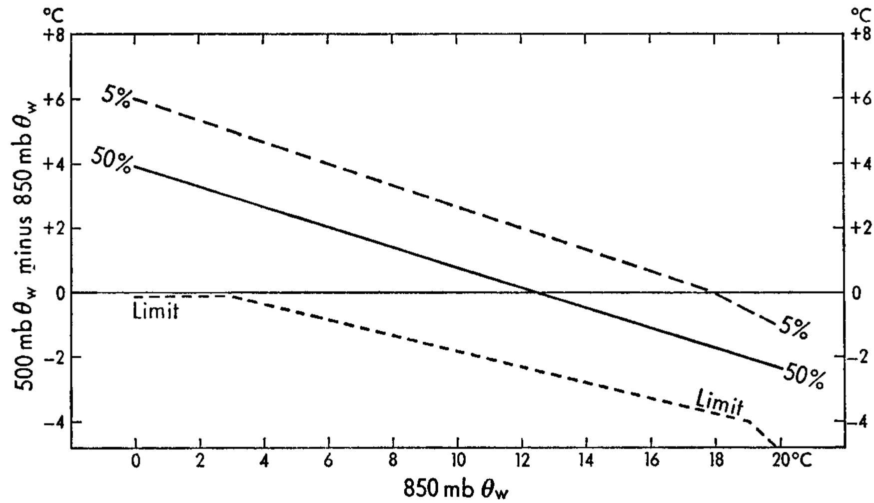

FIGURE 7—RELATIONSHIP BETWEEN 850 mb θ_w AND THE POTENTIAL STABILITY (500 mb θ_w MINUS 850 mb θ_w) ON THUNDERSTORM DAYS

The upper pecked line shows the upper limit of stability which was only equalled or exceeded by 5 per cent of the observations. The lower pecked line shows the most unstable conditions observed during the period. The mean value observed is given by the line marked '50%'.

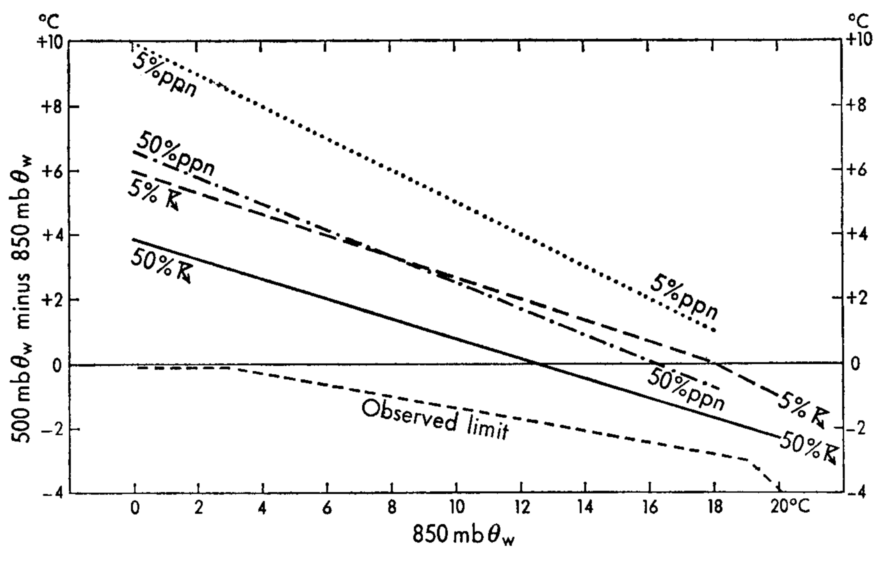

FIGURE 8—RELATIONSHIP BETWEEN 850 mb θ_w AND THE POTENTIAL STABILITY ON THUNDERY AND NON-THUNDERY OCCASIONS

The lines marked '5%' and '50%' with thunderstorm symbols are repeated from Figure 7. The lines marked '5% ppn' and '50% ppn' show the more stable conditions observed when precipitation (other than drizzle) occurred at or very near a radiosonde station at 0000 or 1200 GMT without thunder being reported.

STABILITY INDEX USED IN COMBINATION WITH SYNOPTIC CHARTS

Figures 9 and 10 show a situation when summer thunderstorms occurred both by day and by night. In Figure 9 the isopleths of θ_w have been added to the 850 mb contour chart, together with the position of the surface fronts. It may be seen that θ_w was high over most of Europe. Figure 10 shows the 300 mb contours. SFLOC reports were not used in this case.

Although the air was potentially unstable over much of Europe, the thunderstorms were only observed near and ahead of the 300 mb low and its associated trough. Although England lay just outside the area of potential instability at midnight, the northward advection of warm moist air from France reached Crawley during the morning and the midday ascent showed potential instability.

Figure 11 shows an example of winter thunderstorms. SFLOC reports are represented by lightning symbols. There are three main areas where SFLOC positions appear. One area is over the Atlantic between about longitudes 12° and 22°W which lay beneath the 300 mb low. The second area is associated with the frontal system shown over south-west England and western France. The third area extends from Italy to Greece and the extreme west of Turkey, near and ahead of the 300 mb trough and the cold front preceding it. Almost every SFLOC report falls within the stippled areas where the potential stability index was 3°C or less. Figure 7 suggests that this value is more appropriate for winter thunderstorms than the value of −2°C used for the August example in Figure 10.

CONCLUSIONS

Isopleths of the wet-bulb potential temperature (θ_w) drawn on the 850 mb surface were found to be a valuable aid in the analysis of air-mass characteristics in the lower levels of the troposphere. These isopleths enabled fronts to be located more precisely on small-scale charts which show only a limited number of surface observations.

The value of the 850 mb θ_w was found useful as a guide to the probability of precipitation falling as snow rather than rain. High values of the 850 mb θ_w appeared to be associated with the development of summer thunderstorms over the British Isles and Europe.

Analysis of θ_w at two different levels allows a potential stability index to be derived. The 850 mb and 500 mb levels were found to be the best for use all the year in temperate latitudes. The index of potential stability was obtained by subtracting the 850 mb θ_w from the 500 mb θ_w. The index provided a guide to the risk of thunderstorm development when used in conjunction with routine surface and upper-air charts. The index also showed some value as a guide to non-thundery rain but was of no value when the precipitation fell as snow.

ACKNOWLEDGEMENTS

I am grateful to Mr J. Tate who wrote the program used to compute values of wet-bulb potential temperature, and also to Dr S. R. Mattingly who provided facilities for visual display and subsequent print-out of wet-bulb potential temperatures calculated from current upper-air data stored in the basic analysis data sets of the computer.

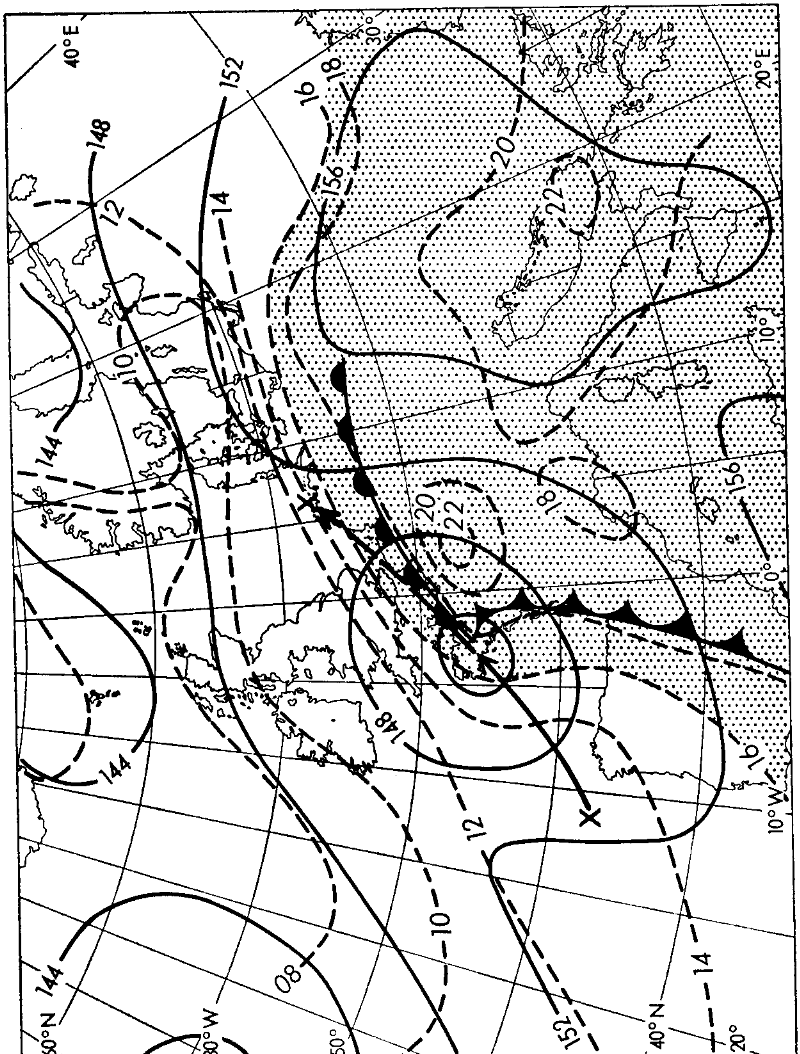

FIGURE 9—CHART FOR 0000 GMT ON 4 AUGUST 1974 SHOWING 850 mb CONTOURS AND ISOPLETHS OF θ_W, WITH POSITIONS OF SURFACE FRONTS ADDED

Areas with values of θ_w of 16°C or more are stippled. Arrows mark the movement of the 850 mb centre during the preceding and subsequent 24 hours.

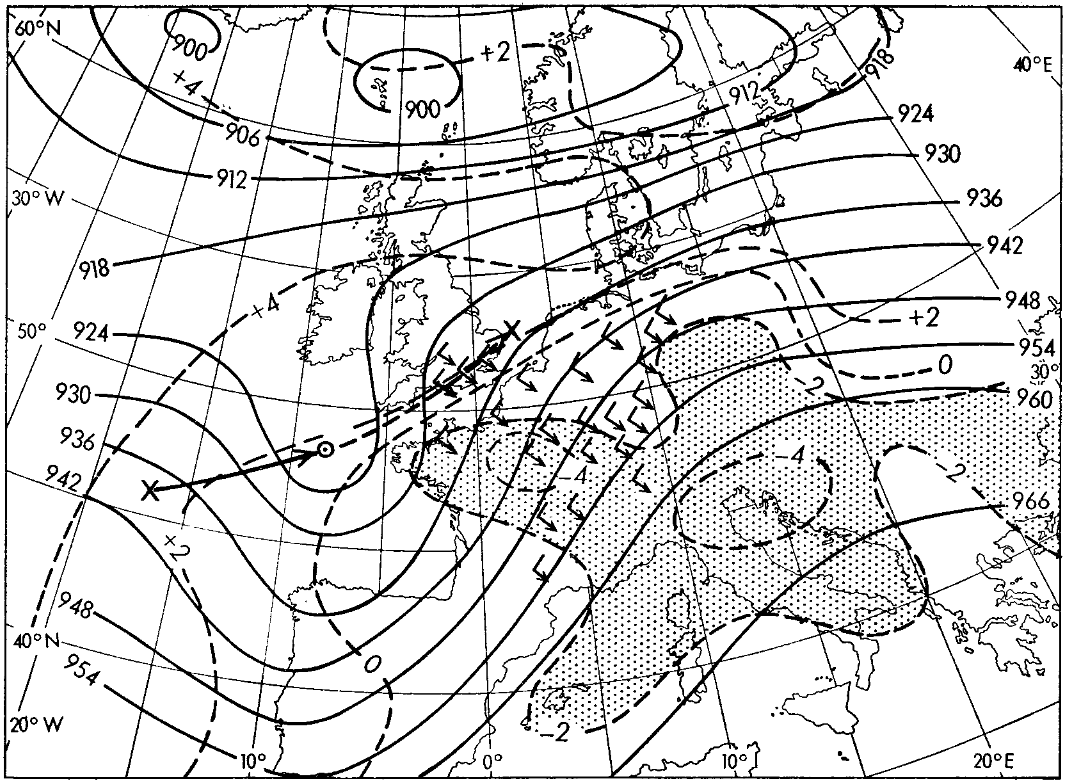

FIGURE 10—CHART FOR 0000 GMT ON 4 AUGUST 1974 SHOWING 300 mb CONTOURS AND ISOPLETHS OF POTENTIAL STABILITY (500 mb θ_W MINUS 850 mb θ_W)

Areas with potential stability values of —2°C and below are stippled. The places at which thunderstorms were reported during the period 0600–1200 GMT are marked with lightning symbols. SFLOC reports were not used in this case. Arrows indicate the movement of the low during the preceding and subsequent 24 hours.

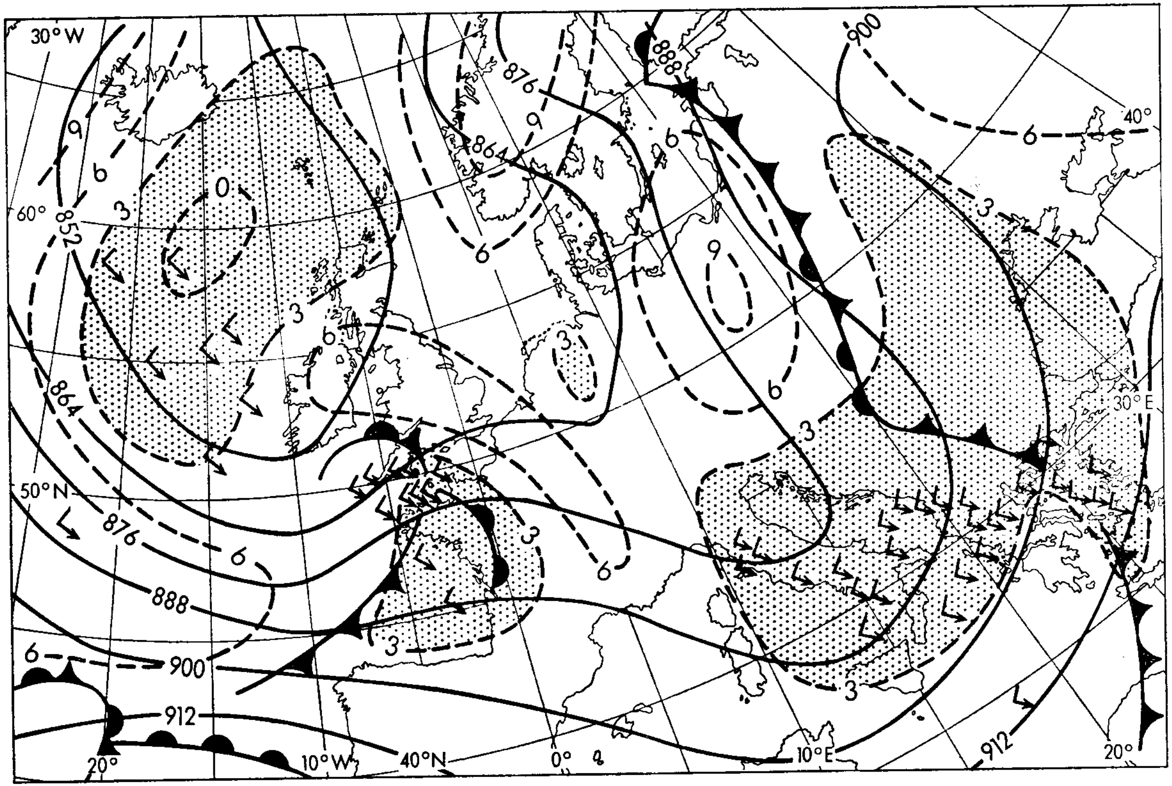

FIGURE 11—CHART FOR 1200 GMT ON 13 JANUARY 1977 SHOWING 300 mb CONTOURS AND ISOPLETHS OF POTENTIAL STABILITY (500 mb θ_w MINUS 850 mb θ_w)

The positions of surface fronts have been added. SFLOC reports for the period 1200–1730 GMT are marked by lightning symbols. The stippled area shows where the potential stability was 3°C or less. This value was found more suitable for use in January than the value of —2°C used in August.

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SOME EXPERIMENTS TO CONTROL THE BUOYANCY OF EXTENSIBLE BALLOONS IN THE STRATOSPHERE

By J. S. FOOT, E. L. SIMMONS and A. E. WHITTAKER

SUMMARY

It is often desirable to control the trajectory of a balloon. Common requirements are: (a) slow ascent or descent to lengthen time of observation, (b) slow ascent or descent to reduce swing of package, (c) floating to give a long period of observation at one altitude, and (d) descent rather than ascent to reduce pollution at the sonde from the balloon. This report looks at some methods that have been used to control extensible balloons.

INTRODUCTION

This report describes some methods which have been used to control the trajectories of extensible balloons in the stratosphere. The design aim was to achieve a slow descent of about 1·5 m s^{-1} from 30 mb to the tropopause. This was required for a pollution experiment being carried out in the stratosphere during daytime. Extensible 2 kg balloons having free lifts between 2 and 4 kg, which gave ascent rates between 4 and 6 m s^{-1} in the troposphere, were used in the study. In this regime, where the flow is turbulent, the ascent rate is not very sensitive to free lift, and different rates can be attributed to different balloon shapes as much as to any other parameter. At lower ascent rates the flow becomes less turbulent, as indicated by the reduction in swing of the system (Foot *et alii*, 1974, and Pettifer and Flavell, 1976), and is very sensitive to the free lift.

The buoyancy is a function of the temperature difference, T_D, between the gas in the balloon and the environment. In particular, the buoyancy is changed by vertical gradients in the stratosphere, by solar heating, ventilation and radiative cooling of the balloon's fabric, and by adiabatic compression or expansion of the gas in the balloon. If T_D remains constant, the balloon will not accelerate, neglecting changes in the drag.

One method of reducing the buoyancy in the stratosphere is to fly two balloons. The first is designed to have near zero free lift after the other balloon has been cut off. The second balloon gives to the combination a strong net free lift. Such an approach has been reported recently by Pettifer and Flavell (1976) where the remaining free lift was +190 g,* as determined from the filling, and the resulting ascent rate was about 1·5 m s^{-1}. The total lift of the gas in the balloon was about 7600 g, and therefore 190 g represents only 2·5 per cent of the total lift. This small proportion is equivalent to a change in T_D of only 5 K in an atmosphere at a temperature of 200 K. The fact that this approach worked suggests that there was some stabilization. One possibility is that there was a balancing between the solar heating, which caused an acceleration upwards, and the ventilation, which reduced T_D as the speed increased. The cooling by expansion would evidently not be fast enough to upset this balance. For a descending balloon, however, although solar heating and adiabatic compression will tend to reduce the velocity, if the balloon descends rapidly enough, the cooling by ventilation will exceed the solar heating and it will accelerate downwards.

In our view, to obtain a floating or slowly descending trajectory, it is necessary to release gas from the balloon in a controlled manner. Provided that there are no sharp temperature changes in the stratosphere, it should be possible to control the speed of descent by controlling this release. Solar heating and adiabatic compression are stabilizing influences.

Experiment and observations

One form of balloon valve consists of a pipe inserted in the balloon neck with a lightly sprung lid on the inside of the balloon. A string from the lid passes across the balloon and is secured at the top. By adjusting the length of the string, the valve can be made to operate at any height. Once open, the free-lift gas is vented out through the pipe. The more constricting the pipe, the longer the time taken to remove the free lift and the higher the balloon rises above its opening altitude. The valve will finally shut between this level and the opening level. It appears that a long narrow pipe shuts after a large negative buoyancy has been gained, whereas a short wide pipe results in almost floating conditions. A valve opening at about 70 mb with the equivalent of 2 kg free lift to dispense requires a pipe about 75 mm long and 48 mm in diameter to give an initial descent rate of 1·5 m s^{-1}. This behaviour has been modelled and the results are shown in Figure 1. Balloons controlled in this manner by a single-action valve do give surprisingly good results, particularly at night-time when there is no solar heating. On one occasion at night a balloon descended initially at 0·4 m s^{-1} but slowed after an hour to 0·25 m s^{-1}. In the daytime the deceleration can be more rapid but its extent is rather variable. For experiments at night or requiring only a limited height range in the daytime, this simple technique may prove useful.

A more complex system has been built and tested which measures the descent rate and opens or closes a valve, depending on whether the speed was slower or faster than a particular rate. The descent rate, $-\mathrm{d}z/\mathrm{d}t$, is given by:

$$-\frac{\mathrm{d}z}{\mathrm{d}t}=-\frac{\mathrm{d}z}{\mathrm{d}p}\times\frac{\mathrm{d}p}{\mathrm{d}t}=\frac{H}{p}\frac{\mathrm{d}p}{\mathrm{d}t},$$

* 'Buoyancy' and 'free lift' are forces, and their appropriate SI unit is therefore the newton. However, they are actually measured by finding suitable weights to balance them, and it is convenient for the practical worker to state their measure in units of mass (grams or kilograms).

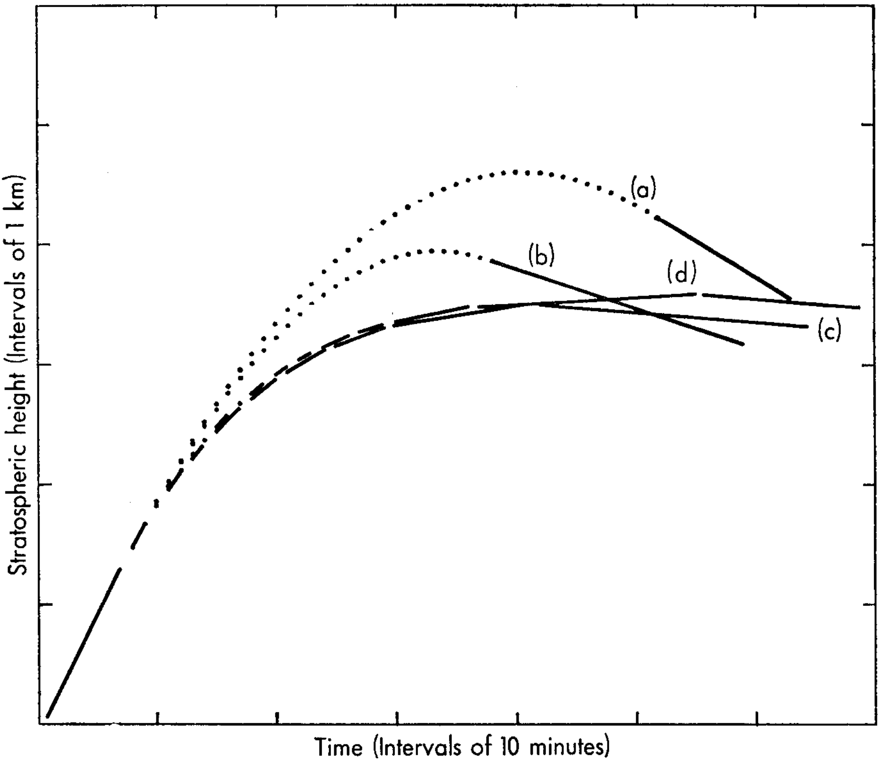

FIGURE 1—COMPUTER SIMULATION OF THE BEHAVIOUR OF A POPPER-TYPE BALLOON NECK VALVE

The initial free lift was assumed to be half the gross load, and the volume discharge rate was assumed constant when the valve was open. Time steps were taken every minute; the continuous lines show where the valve was closed. Curves (a), (b), (c) and (d) refer to rates such that the time taken to lose free lift at the initial opening pressure would be 30, 25, 17 and 10 minutes respectively. The atmosphere was assumed isothermal with a scale height of 6 km.

where p is the pressure at the height z, and H is the scale height ($\approx$ 6 km). Differentiation of the slowly varying analogue output of the pressure sensor is achieved by a pulse-counting technique which avoids the necessity of very low leakage sample-and-hold circuitry. A frequency, f, is generated where $f \propto p^k$, k being a constant, and is counted for two constant equal periods, τ, separated by a time, t, obtained by counting f until it reaches a number N. If these two counts are n_1 and n_2 then:

$$\frac{\mathrm{d}p}{p} = \frac{1}{k}\frac{(n_2 - n_1)}{\tau f}.$$

Hence
$$-\frac{\mathrm{d}z}{\mathrm{d}t} = \frac{H(n_2 - n_1)}{kfT\tau} = \frac{H(n_2 - n_1)}{kN\tau}.$$

The pressure sensor for use with this technique must have low hysteresis, low short-term random variability and low attitude sensitivity. A vibrating wire sensor developed within the Cloud Physics Branch (Whittaker, 1977) has these qualities and, in addition, generates the required frequency directly. An evacuated high-compliance bellows applies tension to a tungsten wire via a pivoted beam. A wire vibrates in a transverse magnetic field, and its vibrations are sensed with a piezo-electric bimorph.* Vibration is maintained by passing a current obtained from the amplified output of the bimorph through the wire. The frequency is proportional to the square root of the pressure, so that $k = 0{\cdot}5$. The values chosen to give a descent rate of $1{\cdot}2$ m s^{-1} are:

$$N = 2^{14},\ \tau = 4\ \text{s, and}\ n_2 - n_1 = 6.$$

The latter condition is detected by feeding the frequency into an 8-bit up/down counter. This is pre-set to an initial count of 5 and counts up for 4 seconds, waits until a second counter has counted 2^{14} of the basic frequency, and then counts down for 4 seconds. The most significant bit of the up/down counter is then 0 if the descent rate is below the desired value, and 1 if it is above. This digital logarithmic differentiator has the advantage that it uses the averaged output of the pressure sensor; assessment of the descent rate is made approximately every 20 seconds and operates a valve in the neck of the balloon pneumatically. Pressure is applied from a reservoir though a solenoid valve to metal bellows in the balloon valve. A constriction between the reservoir and the solenoid valve gives the system an inherent time-constant for opening operations, requiring about 1 minute to open the valve half-way while allowing it to shut instantaneously. This reduces the effect of any error in the pressure differentiating system.

A number of flights of this system and its prototypes have been conducted. Results obtained from a flight from Beaufort Park during the afternoon of 24 May 1976, when the frequency of the pressure sensor and the valve operations were telemetered, are described. The valve was opened when the sonde reached 100 mb and was set to control at a descent rate of $1{\cdot}2$ m s^{-1}. The valve remained shut below 140 mb. Figure 2 shows the descent from near the top of the trajectory. The valve took about 20 minutes to remove the free lift of 3 kg. Also plotted is the descent velocity, the approximate duration and time of the valve operations, and the temperature sounding obtained from the midday Crawley sonde.

The points of interest are:

(a) The valve first shut approximately 45 seconds after the highest altitude.

(b) Many valve operations were made but only the long ones dispensed much gas because of the pneumatic time-constant. These long operations became less frequent as the balloon descended.

(c) The descent rate initially was about $1{\cdot}5$ m s^{-1} but it accelerated to around $1{\cdot}8$ m s^{-1}. These values are larger than the design figure of $1{\cdot}2$ m s^{-1} because the up/down counter had a maximum error of 2, which meant that the balloon would gradually take up the rate defined by a count difference of 8, not 6. This rate is $1{\cdot}6$ m s^{-1}. There was also some noise in the system.

(d) The acceleration at $16{\cdot}0$ km to $1{\cdot}8$ m s^{-1} correlated with the balloon moving into a warmer environment.

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* A bimorph is a strip composed of two layers of oppositely polarized material, arranged so that an electromotive force is produced when the strip is bent.

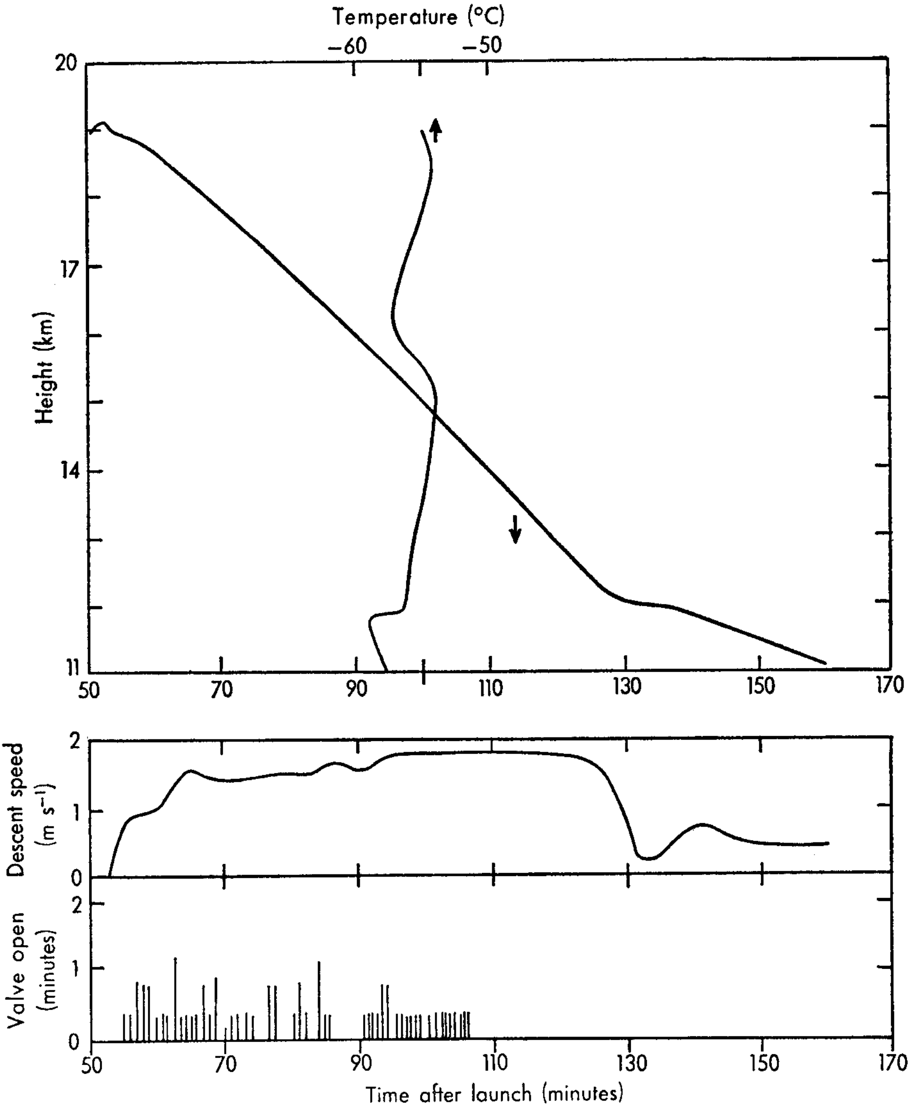

FIGURE 2—TRAJECTORY, TEMPERATURE PROFILE, DESCENT RATE AND VALVE OPERATION FOR CONTROLLED DESCENT

(e) At 14·2 km the valve was disenabled, and from this height down to 12·0 km there was no slowing of the balloon.

(f) At 12 km the balloon almost floated owing to the sudden environment temperature decrease of about 3 K at the tropopause. The Crawley ascent shows this discontinuity at 11·8 km. A change from 214 K to 211 K corresponds to an increase of approximately 1·5 per cent in the buoyancy of the gas in the balloon. The total buoyancy was of the order 6 kg, so that the free-lift change, which almost stopped the balloon, was about 90 g.

(g) In the troposphere, once the effect of the sharp discontinuity had been traversed, the descent rate increased to around 2·3 m s^{-1} in the progressively warmer troposphere.

CONCLUSIONS

It is clear from this work that, to obtain slow rates <2·0 m s^{-1} from an extensible balloon in the stratosphere, the buoyancy required is of the order 100 g. As such, sharp temperature structures of the order of 3 K can completely change the behaviour of slow balloons. Solar heating and ventilation are important in determining the buoyancy, but compression or expansion of the gas in the balloon is generally too slow to be important.

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THE CONTRIBUTION OF EXTRATROPICAL SEVERE STORMS TO THE STRATOSPHERIC WATER-VAPOUR BUDGET

By S. R. MATTINGLY

SUMMARY

By using a climatology of Indian storm tops penetrating into the stratosphere, it has been estimated that severe extratropical storms contribute in the region of 23 per cent as much water vapour to the stratosphere as the Hadley cell circulation. Considering the uncertainties of the calculation, this value agrees reasonably well with values calculated by other authors, and shows that such storms may be a far from negligible source of stratospheric water vapour.

INTRODUCTION

There are a number of possible mechanisms of moisture supply to the stratosphere. These include: (a) upward transport in the Hadley cell circulation in tropical regions, (b) oxidation of methane by excited atomic oxygen and OH molecules in the stratosphere, (c) exchanges of tropospheric and stratospheric air through the tropopause gaps associated with the subtropical and polar jet streams, and (d) direct injection by mixing of stratospheric air with storm tops which penetrate the tropopause.

In mechanism (a) the water vapour must pass through the 'cold trap' of the tropical tropopause and hence most of it is frozen out. Brewer (1949) gives this as the reason why the stratosphere is so dry. This hypothesis is supported by the similarity of the typical value of mass mixing ratio of 3×10^{-6} g/g measured for most of the lower stratosphere (see, for example, Harries, 1976 and Mastenbrook, 1968), and the saturated mixing ratio of 2×10^{-6} g/g at the tropical tropopause. Quantification of this source by Weickmann and Van Valin (1972), hereinafter referred to as WV, using the method of Newell (1970), gives a water-vapour flux of $2{\cdot}2 \times 10^{11}$ kg/year.

Mechanism (b) is not well quantified because global production rates of methane are not particularly well established, but WV give a value of $1{\cdot}1 \times 10^{11}$ kg/year, half the value due to the Hadley cell.

Mechanism (c) is even less well understood. It is evident that very large amounts of air are exchanged between troposphere and stratosphere through the tropopause gaps. Danielsen (1968) calculates that 75 ± 25 per cent of the stratospheric mass is exchanged each year by all processes. Reiter (1975) gives a value of 73 per cent, of which about 20 per cent is due to mechanism (c). It is therefore very difficult to assess the net transport of water vapour by this mechanism. WV assume no net flux occurs, but they indicate that this figure is the difference between a large inflow and a large outflow and hence subject to considerable error. Since Reiter (1975) calculates that mechanism (c) exchanges about half as much mass as mechanism (a), it at least seems unlikely that the former mechanism dominates the latter in respect of water-vapour transport.

This paper attempts to assess the contribution of mechanism (d), which has also been studied by Sissenwine *et alii* (1972), Kuhn *et alii* (1971) and by WV, and is mentioned in Reiter (1975). The methods and results of these authors will be discussed after the description of the method used in this study.

OCCURRENCE AND GENERAL FEATURES OF EXTRATROPICAL STORMS

Many areas in the tropics have upwards of 200 thunderstorm days a year, but the most destructive storms are found outside the tropics in regions near the subtropical jet streams. The main areas affected are: (i) Europe, which has severe thunderstorms on a few days a year (some of these occurring in the British Isles); (ii) the United States of America, which has the greatest observed frequency of tornadoes and large hail in the world; (iii) India, especially in the pre-monsoon season; and (iv) Australia, China, South Africa and Uruguay.

The tops of convective clouds can be divided into two types. In the first, each top is essentially transient and has a lifetime of a few minutes before falling back and being replaced by a similar top. It is believed (Fujita, 1974) that the lifetime of these tops is related to the Brunt–Väisälä frequency, or to the frequency of stable oscillation about the tropopause. In the second type the protrusions are considerably larger in diameter (15 km compared to 5 km) and much longer lived. They occur when the storm becomes well organized and are called 'super cells' (see, for example, Roach, 1967).

In order to obtain information about the nature of storm tops, a study was made of 25 Indian storms measured by Cornford and Spavins (1973). The average linear dimension of the radar echo of transient tops at the tropopause was plotted against the distance of the penetration into the stratosphere by each of the tops. The result is shown in Figure 1. The correlation between the two variables is $+0{\cdot}94$, which is significant beyond the $0{\cdot}1$ per cent level, and the best-fit straight line is:

$$L = (h + 105)/0{\cdot}2719, \quad \ldots \quad (1)$$

where L is the linear dimension at the tropopause in metres, and h the penetration in metres.

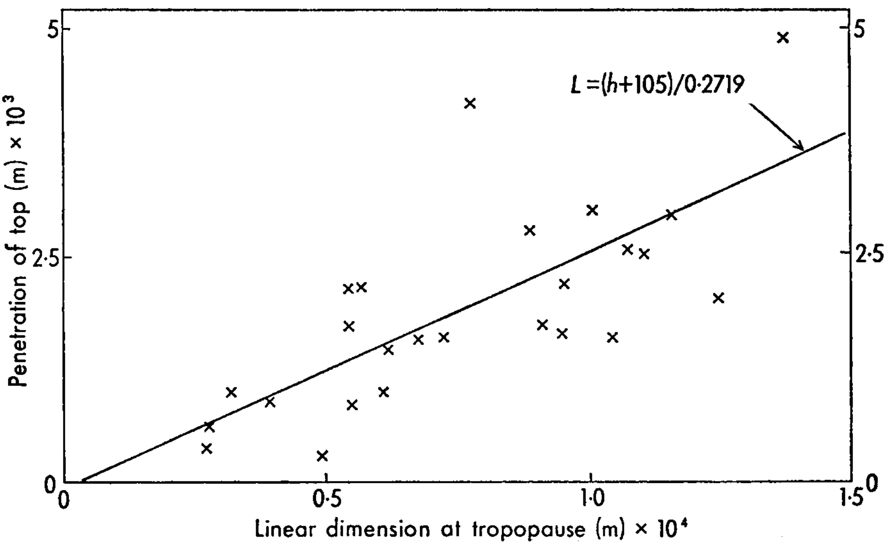

FIGURE 1—PENETRATION THROUGH THE TROPOPAUSE (h) AS A FUNCTION OF LINEAR DIMENSION AT THE TROPOPAUSE (L)

It is interesting to note that one of the storms, which was not included in the data despite having a linear dimension of $3{\cdot}0 \times 10^4$ m at the tropopause, only penetrated some 4000 m into the stratosphere. It seems reasonable to assume that this top is a super cell and has reached the maximum penetration possible, while all the other tops are transient and have not attained their maximum penetration.

Long *et alii* (1965) state that the American radar echoes of tops penetrating into the stratosphere have a base diameter to height ratio of 5:1. Lee and McPherson (1971) state that Malaysian tops are somewhat thinner. From equation (1) it appears that Indian tops are also somewhat thinner than the American ones.

CALCULATION OF WATER-VAPOUR INPUT TO THE STRATOSPHERE

As the detailed physics of the penetration process is not well understood, it was decided to rely on measurements of water-vapour column density upwind and downwind of penetrating tops. Such measurements have been made by Kuhn *et alii* (1973) and by Barrett *et alii* (1972). A total of five tops which penetrated well into the stratosphere have been measured.

The input of water vapour from a penetrating cloud top was modelled by assuming that, as the air travelled through the top, it picked up water vapour by mixing and evaporation of entrained ice crystals, and that the amount of water vapour picked up was the same over the whole area of the cloud. This leads to an expression for the rate of water-vapour injection, W:

$$W = hLV\rho \text{ g s}^{-1}, \quad \ldots \quad \ldots \quad \ldots \quad \ldots \quad (2)$$

where h is the penetration in metres of the top through the tropopause, L is the linear dimension of the top at the tropopause in metres, V is the stratospheric wind speed relative to the top in metres per second, and ρ is the mass of water vapour per unit volume picked up by the air in its passage through the top. In this study V was assumed to be 10 m s^{-1}, which was appropriate for the rather low wind shears measured by Cornford and Spavins (1973) in India. The value of ρ was calculated from the data of Kuhn *et alii* (1973) and Barrett *et alii* (1972) to be $2{\cdot}4 \pm 0{\cdot}8 \times 10^{-9}$ g cm^{-3}.

Instead of attempting to estimate the lifetime of each top, radar climatologies, which measure (at the same time each hour) the number of tops which are penetrating the tropopause to varying degrees, were employed and the measured number of tops was assumed to last for the ensuing hour. Thus, if n tops are seen by radar over a given area to be penetrating the tropopause by a distance h, we have that the water vapour injected into the stratosphere in that area during the ensuing hour will be:

$$I = 3600nhLV\rho \text{ g.} \qquad \ldots \quad \ldots \quad \ldots \qquad (3)$$

Using the assumed values of V and ρ, and the relation between h and L (equation (1)), gives:

$$I = 318nh(h + 105) \text{ g.} \qquad \ldots \quad \ldots \quad \ldots \qquad (4)$$

CLIMATOLOGY OF INDIAN STORM-TOP PENETRATIONS

Kantor and Grantham (1968) give an extensive climatology of American storm-top heights and these data were employed by Sissenwine *et alii* (1972) in their study of water-vapour injection into the stratosphere. In order to obtain an independent assessment, the unpublished climatology of Calcutta storm tops by D. K. Rakshit (personal communication to S. G. Cornford) was used here. This climatology is preferred to that of Bhattacharyya and De (1966) since the latter measurements are believed to underestimate the heights reached (Cornford, personal communication). The number of storm tops, measured on an hourly basis, which penetrate the tropopause in a year by a given amount are shown in Table I.

TABLE I—THE NUMBER OF STORM TOPS PER YEAR WHICH PENETRATE A GIVEN AMOUNT INTO THE STRATOSPHERE AT DUM-DUM, CALCUTTA

Penetration (m)	Number per year
0–610	19
610–1220	17·5
1220–1830	15·5
1830–2440	14
2440–3050	11
3050–3660	7·5
3660–4270	2
	Total 86·5

From equation (4) we can thus calculate the water vapour injected into the stratosphere by storm tops which penetrate to varying degrees through the tropopause. The results are collected in Table II.

TABLE II—WATER-VAPOUR INJECTION INTO THE STRATOSPHERE IN A YEAR BY STORMS WHICH PENETRATE A GIVEN AMOUNT THROUGH THE TROPOPAUSE

Penetration (m)	Water-vapour injection (kg)
0–610	$0{\cdot}8 \times 10^6$
610–1220	$5{\cdot}2 \times 10^6$
1220–1830	12×10^6
1830–2440	21×10^6
2440–3050	27×10^6
3050–3660	28×10^6
3660–4270	10×10^6
Total	$1{\cdot}0 \times 10^8$

It is interesting to note that a considerable proportion of this injection is contributed by the very high tops, and this water vapour is more likely to remain in the stratosphere than that injected into the transition layer between troposphere and stratosphere which can be many hundreds of metres thick.

Figure 13.1 on page 391 of Palmén and Newton (1969) shows those areas of the world which are subject to severe storms, and the total area outside the tropics can thus be roughly computed. The area which is observed by the Dum-Dum radar is about 0·2 per cent of the total.

Therefore we estimate, by this method, that severe extratropical storms contribute $1{\cdot}0 \times 10^8 \times 100/0{\cdot}2 = 5{\cdot}0 \times 10^{10}$ kg/year, or 23 per cent as much water vapour to the stratosphere as the estimate of WV for the Hadley cell.

This is admittedly a rough estimate but it suggests that the direct injection of water vapour into the stratosphere by convective clouds is an important source of stratospheric moisture.

COMPARISON WITH OTHER WORKERS

WV assume that a 100 m thick layer of the storm anvil mixes with the stratospheric air. They take as a representative sample a storm studied by Fujita (1972) with an anvil area of 60 000 km². They then estimate the effect of 100 storms per day of this size around the globe penetrating into the stratosphere. This gives a global water-vapour flux of $8{\cdot}8 \times 10^{11}$ kg/year, or four times their Hadley cell flux.

Sissenwine *et alii* (1972) give a detailed climatology of tropopause penetrations over the United States, based on the work of Kantor and Grantham (1968). They also consider the shape of the penetrating tops, and hence calculate the amount of tropospheric air taken up into the stratosphere annually. This value is then converted to a global value on an area basis. Finally, they assume that 1 per cent of this injected water vapour is released into the stratosphere. This gives the result that the storms inject 70 per cent of the WV Hadley cell flux.

Reiter (1975) estimates that small-scale disturbances, such as tropical storms, exchange annually a small amount of air between stratosphere and troposphere, perhaps nearer 1 per cent than 5 per cent of the stratospheric mass, compared to the 38 per cent exchange by the Hadley cell circulation. Sissenwine *et alii* (1972) give a value of ½ per cent. It is because this air is so rich in moisture that the contribution to the water-vapour budget is relatively large.

The 36 500 storms per year assumed to penetrate the stratosphere by WV agrees fairly well with the 67 000 storms calculated by Sissenwine *et alii* (1972) and the 43 000 estimated in the present paper.

The injection of water vapour by an average storm is given by WV as 7×10^7 kg, and this is considerably larger than the $1{\cdot}2 \times 10^6$ kg estimated in the present paper and the $2{\cdot}2 \times 10^6$ kg assumed by Sissenwine *et alii* (1972). Kuhn *et alii* (1971) give a value of $2{\cdot}0 \times 10^5$ kg for Oklahoma storms by assuming that each cell has an area of 100 km^2 and injects 20×10^{-4} g cm^{-2} of water vapour. This seems to be a minimum value as they assume no downstream spread of the water vapour.

Using the method described in the present paper for calculating water-vapour injection, together with the extensive climatology of Kantor and Grantham (1968) gives an estimate of 30 per cent of the WV Hadley cell flux. This represents, in the author's opinion, the least arbitrary estimate of the contribution of severe storms to the stratospheric water-vapour budget, although it is still a far from satisfactory one.

Conclusion

Using a climatology of Indian storm tops penetrating into the stratosphere, it has been estimated that severe extratropical storms contribute in the region of 23 per cent as much water vapour to the stratosphere as the Hadley cell circulation. This value, considering the uncertainties of the calculation, agrees reasonably well with values calculated by other authors, and shows that such storms may be a far from negligible source of stratospheric water vapour.

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NOTES AND NEWS

Seychelles joins WMO

The Seychelles became a Member of the World Meteorological Organization on 17 March 1977 and will, in consequence, take over the UK seat on the WMO RA I Tropical Cyclone Committee for the South-west Indian Ocean.

Meteorological services for civil aviation in Cyprus

The remaining UK responsibilities for meteorological services for civil aviation in the Republic of Cyprus ceased on 4 January 1977 when the Cyprus Meteorological Service took over responsibility for Sigmet Warnings in the Nicosia Flight Information Region.

The meteorological office at Salālah

Meteorological Office staff at the observing office at Salālah in the Sultanate of Oman were withdrawn on 29 March 1977. From that date the office was taken over as a going concern by Pan American Airways on behalf of the Oman Government and is being run as a forecast office for the developing civil airport.

New posts in agricultural meteorology

Three new scientific posts in agricultural meteorology have been agreed, with effect from April 1977, for work with the Agricultural Development and Advisory Service of the Ministry of Agriculture, Fisheries and Food. They are at Reading, Cambridge and Harrogate.

Improved system for estimation of evaporation and soil moisture deficits

The scientific basis of an improved operational system for estimating evaporation and soil moisture deficits was discussed in a paper presented at a meeting of the Meteorological Research Committee on 31 January 1977. The operational system is undergoing trials and is expected to replace the existing Soil Moisture Deficit Bulletin which is primarily of value to agriculture and the water industry.

Warnings of synoptic developments for the Central Electricity Generating Board

Arrangements have been made for the Central Electricity Generating Board to be provided during the afternoon in winter, when commercially important decisions have to be made, with advance warnings of hitherto unexpected developments in the synoptic situation. These warnings will be issued by the Central Forecasting Office and relayed through the London Weather Centre.

Stratospheric air sampler

Development of a balloon-borne device for obtaining samples of air in the stratosphere have begun, and analysis of the samples is expected to provide data on the concentration of several species of importance to the photochemistry of the upper atmosphere. Emphasis in the design is being placed on simplicity, low weight and cost so that existing radiosonde facilities may be used. A prototype device has been built and is under test.

GATE data from the Meteorological Research Flight

By the end of March 1977 all 40 flights carried out by the MRF Hercules aircraft during GATE (the international co-operative project for the Atlantic Tropical Experiment), held in 1974 and based at Dakar in Senegal, had been fully processed. The magnitude of this task can be appreciated from the fact that for each hour of flight 2·3 million words of data were collected, processed and recorded on magnetic tape, and that the 40 missions flown resulted in more than 350 flying hours. The data have been distributed in their completed form to the two World Data Centres at Moscow and Washington, and also to those sub-groups which have direct interest in the parameters measured.

Low-level measurements of turbulence over the sea

Over the past two years a series of flights has been conducted with the Hercules aircraft of the Meteorological Research Flight at low levels (30 m to 1000 m) over the sea to measure turbulence in a variety of meteorological conditions.

The flights in weakly suppressed conditions with little or no cloud have shown that there were marked differences between the turbulence spectra measured on runs along and across the wind, the latter exhibiting structure at wavelengths considerably greater than those measured along wind.

Reception of VHRR pictures at Lasham

The satellite receiving station at Lasham is now equipped to receive segmented Very High Resolution Radiometer (VHRR) pictures from NOAA 5. These are supplied to the Central Forecasting Office and to the extended network of out-stations which receive special satellite facsimile transmissions; these stations are also supplied with low-resolution scanning radiometer pictures. VHRR images are restricted to about 1000 $\times$ 1400 km segments because of the inability of Post Office lines to pass the large amount of information available from VHRR instruments.

European Centre for Medium-range Weather Forecasts

The Property Services Agency (PSA) of the Department of the Environment is building the European Centre for Medium-range Weather Forecasts at Shinfield Park, Reading. The complex, designed in PSA's Southern Region Office, will comprise a computer hall, administration building, and conference and teaching accommodation. On completion in November 1978 it will house the 100 staff employed by the 16 member nations who are at present housed temporarily in Bracknell. Sponsored by the Ministry of Defence, construction is being carried out by John Laing Construction Ltd at a cost of just under £2 million.

The computer hall and supporting work space will total approximately 2000 square metres, and advanced methods of air-conditioning will be used. Stand-by generators, switching and transformers are to be located in a separate building.

The administration building includes the director's suite, library, offices and boiler-house. The conference block consists of main entrance, class-room, dining-room and kitchen, conference room and lecture hall. Both the conference room and lecture hall will be provided with translation and projection facilities, and will be ventilated from plant at roof level.

The site is located next to the existing Meteorological Office College at Shinfield Park. Care has been taken to retain specimen trees on the site and as many as possible of the other trees around the boundary. (See *Meteorological Magazine*, **103**, (1974) p. 28.)

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CONTENTS

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