Conditions for the occurrence of severe local storms

By T. N. CARLSON[1] and F. H. LUDLAM, *Imperial College, London*

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ABSTRACT

The technique of relative-flow analysis on isentropic surfaces is used to examine the large- ("synoptic"-) scale situations associated with selected severe local storms near southern England and over the mid-western U.S.A. (including the Horsham, Wokingham, and Geary storms whose behaviour has been described in several previous publications). The storms occur ahead of major troughs, in the vicinity of confluence-lines (usually recognised as cold fronts over western Europe but as "dry-lines" over the U.S.A.), where an increase of wind with height favours the organisation and intensification of cumulonimbus convection. Extreme instability arises where small-scale convection is confined to a lowermost 1 or 2 km (leading to an abnormally high wet-bulb potential temperature) beneath a plume of very warm air lying downwind of an extensive arid plateau (Spain or Mexico). The instability is released where the (backed) low-level flow eventually reaches the edge of the restraining plume aloft. It appears that the occurrence of severe local storms demands a peculiarly favourable combination of geographical features and atmospheric flow-pattern.

1. Introduction

In a previous paper (Ludlam, 1966) atmospheric convection has been classified according to scale, and the stratifications characteristic of layers occupied by cumulus convection and by cumulonimbus convection have been discussed. Cumulus normally occupy only the lower troposphere and have small buoyancy, but after shower formation in large cumulus with supercooled tops buoyancy is substantially increased by the *precipitation* and *glaciation* of cloud water, and the cumulonimbus which develop reach much higher levels. In the presence of a favourable wind shear the cumulonimbus convection becomes more organised, and a proportion of the updraught air ascends more nearly adiabatically and acquires greater buoyancy, so that updraught-speeds of 20 m sec^{-1} or more are attained. If, in addition, the cumulonimbus convection begins in an atmosphere which has been conditioned by recent deep *dry* convection, rather than *cumulus* convection, and which therefore has unusually steep lapse-rates in the lower troposphere, the cloud buoyancy and updraught-speeds become exceptionally great with the production of the giant hailstones and other violent phenomena of the *severe local storm*. In the present paper we describe the conditions which led to the occurrence of some storms of this kind in western Europe and the middle-western U.S.A.

[1] Now at the National Hurricane Research Laboratory, Miami, Florida.

2. The technique of analysis

(a) *Use of potential temperatures*

We shall frequently refer to the *potential temperature* θ, the *wet-bulb potential temperature* θ_w, and the *saturation potential* temperature θ_s, defined as the value of θ_w corresponding to the state of saturation at an observed temperature and pressure. These reference parameters have a familiar convenience in the discussion of dry and cloudy convection; in particular, a comparison of an average value of θ_w in air near the ground with the values of θ_s at higher levels gives an immediate indication of the sign and magnitude of the temperature-excess which such air may attain there by adiabatic ascent.

Outside the tropics the state of the upper troposphere is governed by large-scale ("slope") convection. Especially in the winter, when the ocean rather than the land surface is the dominant heat source, the small-scale convection occurs principally in the descending

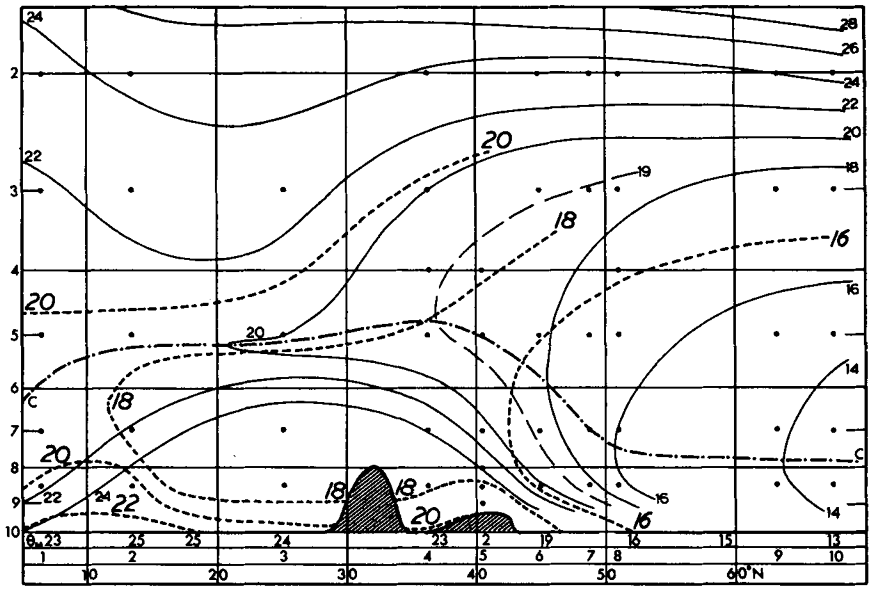


Fig. 1. Mean summer vertical section overland approximately along the Greenwich meridian, July 1959. Thin solid lines show the distribution of θ_s at intervals of 2°C; thicker pecked lines the distribution of θ_w (labelled in larger, italic figures). The dot-dash line marked *C* shows the mean position of the top of the layer which may be occupied by cumulus clouds, above which θ_s increases with height. Height is represented on a logarithmic scale of pressure, the figures on the scale on the left indicating pressure in hundreds of mb.

The small circles indicate the positions of observations used in the analysis, from soundings given reference numbers in the middle panel of the figure base (1: Lagos, 65201; 2: Niamey, 61052; 3: Fort Trinquet, 61401; 4: Gibraltar, 08495; 5: Zaragoza, 08159; 6: Bordeaux, 07510; 7: Trappes, 07145; 8: Crawley 03774; 9: Östersund, 02062; 10: Sodankyla, 02836). The Atlas mountains, on the northern edge of the Sahara, and Spain are shown schematically.

In the upper panel at the base of the figure are entered mean values of screen-level wet-bulb potential temperature in mid-afternoon, in °C.

branches of the large-scale circulations, which are directed towards the equator. There is a progressive rise of θ_w as the air moves into lower latitudes, and in the *ascending* branches the modified air is led poleward into the upper troposphere with values of θ_w or θ_s substantially greater than those prevailing directly beneath. The mean stratification thus produced is unfavourable for cumulonimbus convection.

Even in the summer, when the principal heat sources in the northern hemisphere become land masses, and the tropical cumulonimbus convection and its own nearly-neutral stratification tend to advance into higher latitudes, large-scale convection persists and strongly influences the occurrence and intensity of the cumulonimbus convection. The development of intense cumulonimbus is favoured on the forward side of large-scale troughs, less by a lowering of θ_s aloft than by the provision of wind shears favourable for the organisation of the cumulonimbus, and the local production of an abnormally high θ_w near the ground. The modification of the mean stratification (Fig. 1) is illustrated by the meridional cross-section of Fig. 2 in advance of a trough approaching the west coasts of Europe and North Africa. The distribution of θ_s hardly differs from the mean, but over southern France θ_w near the surface reaches the extraordinarily high value of 24°C, ordinarily found on this section only south of the intertropical front over West Africa. This abnormality was associated with the production of intense cumulonimbus which travelled across England in a frontal zone of strong wind shear (section 4*b*).

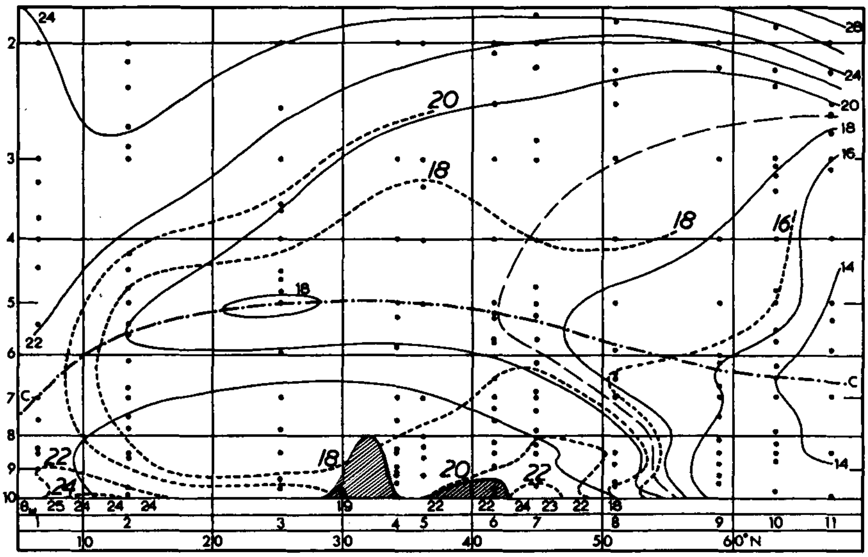


Fig. 2. Vertical section over land, approximately along Greenwich meridian, in advance of a large-scale trough, 9 July 1959. For general explanation see legend to Fig. 1. (Soundings used from 1: Lagos, 65201; 2: Niamey, 61052; 3: Fort Trinquet, 61401; 4: Kenitra, 60119; 5: Gibraltar, 08495; 6: Zaragoza, 08159; 7: Bordeaux, 07510; 8: Crawley, 03774; 9: Stavanger, 01415; 10: Östersund, 02062; 11: Sodankyla, 02836.)

(b) *Use of isentropic charts*

Over most of a large-scale motion system, the flow can be regarded as nearly adiabatic. Only in the rather shallow layer occupied by the small-scale motion must the flow be regarded as essentially non-adiabatic. For the discussion of the properties of the air-streams which arrive in the vicinity of intense cumulonimbus and there determine the stratification of the atmosphere, it proves convenient to use the dry adiabatic reference process and examine the large-scale flow in surfaces of constant θ (that is, to use the technique of isentropic analysis introduced by Rossby (1937) and already shown to be useful in this and other problems). In contrast to the conventional representation in isobaric maps, the flow in isentropic maps contains a direct indication of the vertical component of the motion, which is sufficiently correct to justify the construction of trajectories over periods of up to about 2 days. This proves a great advantage when there is no rapid development or decay of the system and it can be assumed that over such a period it is in a steady state (Green, Ludlam & McIlveen, 1966), for then the streamlines of the flow relative to the system are trajectories, otherwise to be found only by tedious extrapolation using maps for a succession of observation times. We have made analyses on isentropic surfaces with the steady-state assumption and found them to be more illuminating than conventional analyses.

3. The distribution of dry- and wet-bulb potential temperature near the ground

The buoyancy which may develop within cumulonimbus is proportional to the difference between θ_s aloft and the (higher) average value of θ_w in the adiabatic layer produced by small-scale convection, which overland is usually 1–2°C below that recorded at screen-level in the afternoon.

In North America the severe storms occur where the low-level air arrives after being subject to small-scale convection for several days over the southern part of the North Atlantic and the Caribbean Sea. In summer it enters the southern U.S.A. with the rather high mean θ_w of about 22°C in the lowest km. In contrast, near western Europe the sea surfaces are relatively cool, and the increase of the mean θ_w in the layer of small-scale convection which occurs there over land as a result of

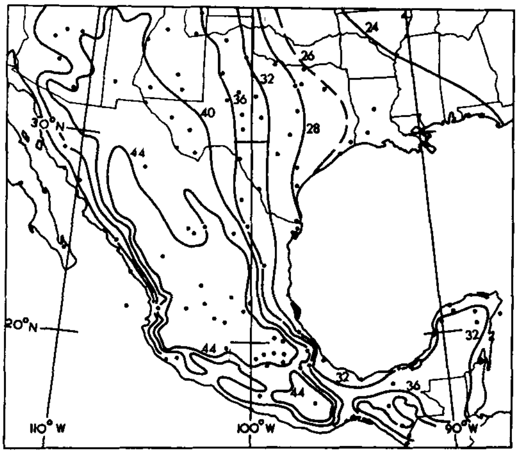


Fig. 3. Distribution of mean daily maximum screen-level potential temperature (in °C) over Mexico and adjoining territory during late April and early May. (Values derived from average of April and May mean data, obtained from stations reporting for more than 3 years in available Mexican climatological data, and from U.S. Weather Bureau information for 1961. Dots indicate locations of stations.)

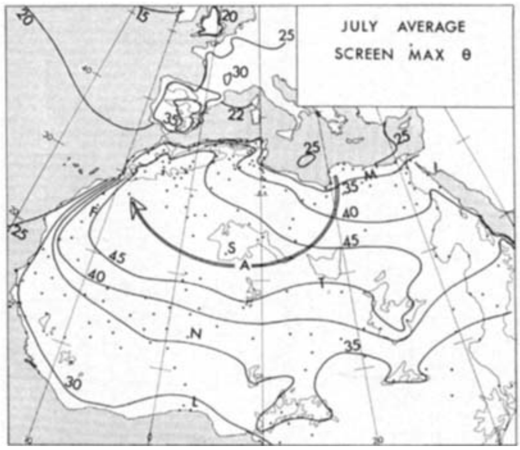


Fig. 4. Distribution of mean daily maximum screen-level potential temperature (in °C) over North Africa and western Europe during summer. (Values obtained from 30-year climatological means for July. Dots show the location of reporting stations.) The diagram includes a streamline over the Sahara for the mean flow in the layer between 850 and 700 mb during July 1959.

sunshine for one or sometimes two days is important in the explanation of the occurrence or severity of the thunderstorms.

The probable increase can be estimated by considering the energy balance of the convection layer, in which the principal item is the net radiation at the ground. Some of this is transferred into the ground, and the remainder into the convection layer, partly as sensible heat and partly as the latent heat of water vapour introduced by evaporation at the surface. It appears (see, e.g., Penman, 1956, p. 18, and Pruitt & Aston, 1963, p. 74) that the former is a rather small fraction of the latter over moist ground, whereas over parched soil it is the reverse (Dyer, 1961; Halstead, 1954, p. 358). If the evaporation is diminished the ground temperature rises and the increase of θ_w in the adiabatic layer is reduced because it becomes deeper.

From the empirical formulae of Monteith & Szeicz (1961) it can be inferred that in cloudless mid-summer weather in the regions which concern us the mean θ_w in a convection layer 150 mb deep should rise by about 2.5°C every 24 hours; in the considerably deeper convective layers which are characteristic of arid regions the increase of θ_w should be substantially less, falling to only about 0.5°C for a depth of 400 to 500 mb. Thus over regions such as Mexico, North Africa and central Spain the layers of small scale convection are deep and dry, characterised by rather low values of θ_w, but high values of θ corresponding to high afternoon surface temperatures (Figs. 3 and 4).

Fig. 4 shows that in July air persistently reaches the central and western Sahara from the Mediterranean, and after several days passage over the desert acquires a θ of about 43°C. The adiabatic layer becomes very deep, extending above the 600 mb level, so that the radiative heat loss becomes comparable to the energy introduced from the surface and little further change in potential temperature can occur. Accordingly a rather uniform Saharan air mass arises, characterised by a θ of about 43°C and a θ_w which is just over 18°C near the surface and 17°C at 600 mb.

Fig. 4 also shows that θ at screen-level in the afternoon, and therefore in the adiabatic layer, also reaches characteristic rather uniform values over smaller regions, depending upon latitude and topography. In particular, over the table-land of Spain (elevation about 1 km) it is about 36°C, while over the moister ground of France it is only 26–28°C. An important consequence is that in a southerly airstream a "plume" of potentially very warm air from Spain is found aloft over southwest France, and acts like a "lid" to confine the small-scale convection there to a layer only 1 or 2 km deep. The rise of θ_w in this layer during a sunny day may therefore become abnormally large, favouring the development of intense cumulonimbus if some mechanism of releasing the deep convection can be found. We shall show that the warm plume is an important factor in the production of severe storms not only in this region, but also in the mid-western states of the U.S.A. Its presence down-wind of its source region is often manifest by the castellanus clouds which appear when condensation commences in it following ascent in the large-scale flow. These clouds are notorious precursors of severe thunderstorms.

4. Case-studies of severe storm situations over western Europe

Detailed studies have been made of three situations leading to severe storms over southern England and Belgium, and of another apparently favourable situation in which no cumulonimbus developed.

(a) *The Horsham storm of 5 September*, 1958

This severe storm formed in the English Channel during the afternoon and passed south of London into the North Sea, there dying out at about midnight.

An almost stationary and slowly filling depression was centred west of Ireland; its cold occlusion entered England from the southwest early on the 3rd, accompanied by a narrow belt of rain with local thunderstorms. On the 4th front became quasi-stationary across Central England, eastern and southern France, and Spain; it became poorly defined and by the morning of the 5th had disappeared from official weather analyses.

The flow in the upper troposphere. The flow pattern was practically stationary, so that the streamlines on the isentropic relative-flow charts are on this occasion trajectories in their correct position relative to the ground.

On the surface for $\theta = 52°C$ (Fig. 5) the front is characteristically seen as the confluence of two distinct air streams, one of which has turned northward after travelling across the southern Sahara as an easterly current, and the other of which arrives from the west after travelling from the rear of the cyclone west of Ireland. The "limiting streamline" which defines the edge of the flow from the Sahara seems to originate south of the intertropical front over the southern Sahara, where θ_w was about 19°C.

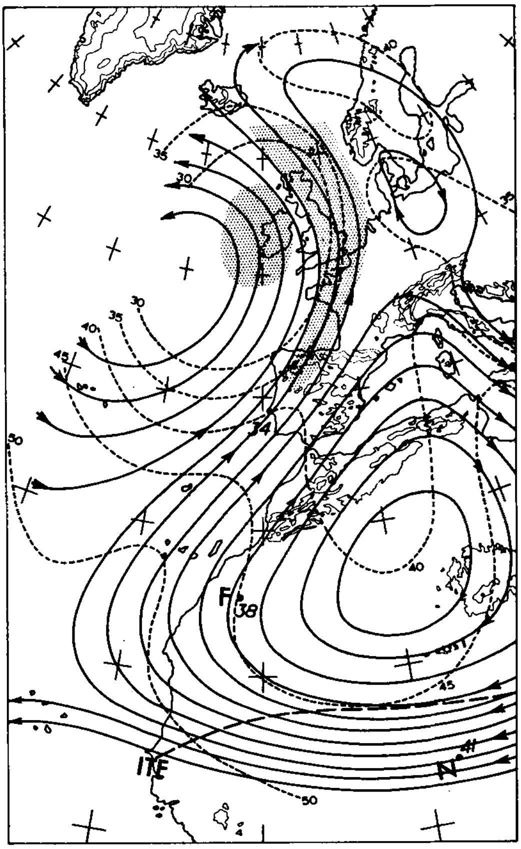


Fig. 5. Isentropic relative-flow chart for $\theta = 52°C$, 1200 G.M.T., 5 September 1958, containing streamlines (solid) and isobars (pecked lines labelled in tens of mb). The pecked line marked ITF is the surface position of the intertropical front, and the letters *N* and *F* mark the positions of Niamey and Fort Trinquet. Italic figures give the adiabatic lifting condensation level (in tens of mb) at these places and Lisbon, and the stippling shows areas of extensive high cloud.

The soundings from Niamey and Fort Trinquet show that where the flow turns northward near the coast of West Africa the condensation level of the air is probably somewhat above the 400 mb level, consistent with the appearance over northwest Spain (where the air has risen to between about 380 and 350 mb) of a belt of cirrus which extends northward into

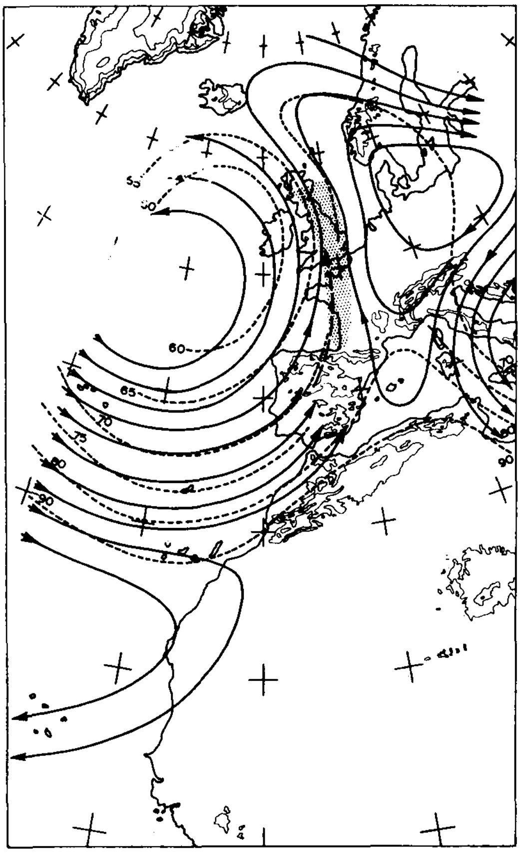


Fig. 6. Isentropic relative-flow chart for $\theta = 33°C$, 1200 G.M.T., 5 September 1958. During the flow over Spain and into the confluence zone farther north, ascent and condensation remove air to higher isentropic surfaces, especially in the belt of extensive altocumulus (with castellanus; stippled area, terminated arbitrarily over Scotland).

England, with a sharply defined western edge along the limiting streamline. Near the front over England θ_s in the high troposphere is limited to the value of 19°C set by the value of θ_w in the flow over Africa.

On the charts for rather lower isentropic surfaces the limiting streamlines of the southerly flow leave Africa in positions farther north, and the air leaving the uppermost part of the adiabatic layer over the Sahara has the significantly lower θ_w of about 17°C, and a condensation level between 480 and 450 mb which is barely attained over southeast England. Consequently this is also about the value of θ_s found there at these levels.

The flow in the lower troposphere. On surfaces of θ less than 43°C the front is the confluence-line between the westerly maritime air flow and a northeasterly flow which enters the Mediterranean from S.E. Europe before becoming an easterly and then a southerly stream. Over land near the confluence-line small-scale convection modifies the streams and complicates their structure.

Fig. 6 shows the flow in the surface for $\theta = 33°C$. On this surface the maritime stream, which has been moistened by small-scale convection over the ocean, crosses Portugal and northwest Spain without substantial modification, but over eastern Spain θ at screen-level reached 35°C on the afternoons of the 4th (Fig. 7) and 5th (Fig. 8), and much of the air which enters this region from either east or west disappears from this isentropic surface. A rather narrow plume extends northwards

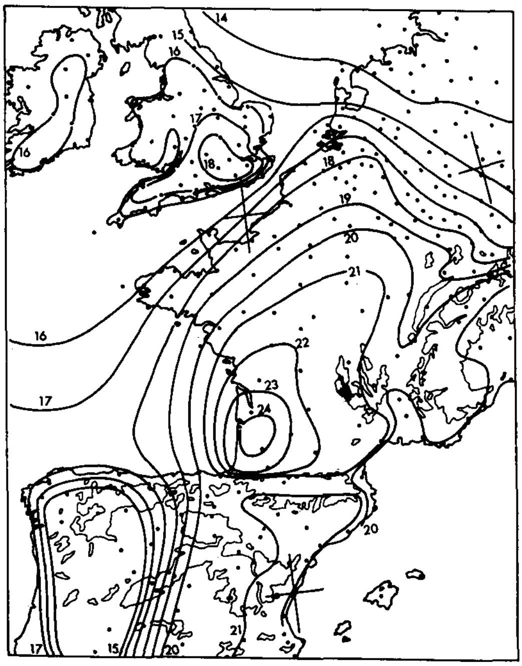


Fig. 7. Distribution of screen-level wet-bulb potential temperature (in °C) over western Europe, 1500 G.M.T., 4 September 1958. Ground height contours for 3000 ft are drawn as thin lines. Dots show the positions of observing stations. At this time the temperatures were about the maxima for the day. The highest wet-bulb potential temperatures are over France in the lee of the Pyrenees, beneath the potentially warm plume from Spain.

marking the base of the layer modified by convection over eastern Spain, and shrinks in width as the air rises and produces castellanus clouds with bases at about the 700 mb level. The sounding at Chateauroux (Fig. 9), on the right of the plume, has a layer with a nearly dry-adiabatic lapse-rate (θ about 36°C) from 850 to nearly 700 mb, above which level the steep lapse-rate and high relative humidity suggest that the castellanus reach to at least 600 mb and perhaps locally above 500 mb. This and other soundings do not show a state of saturation at the base of the castellanus layer, perhaps because of unresponsive humidity sensors but more probably because the clouds usually occur in narrow moist plumes produced by intermediate-scale circulations in the source region (castellanus are well known to appear in

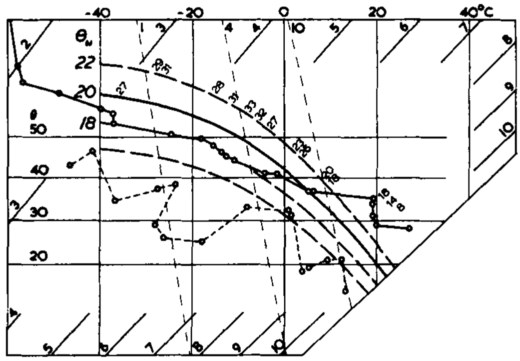


Fig. 9. Sounding for Chateauroux (07354), 1200 G.M.T., 5 September 1958, represented on a tephigram. On this and similar subsequent diagrams isobars are labelled in hundreds of mb, and lines of constant potential temperature θ and wet-bulb potential temperature θ_w are labelled in °C. Thin pecked lines of constant saturation mixing ratio for 1, 4 and 10 g kg^{-1}; are included. This sounding shows air from the Spanish adiabatic layer present between 850 and 700 mb. The "lid" can be considered to be at 850 mb. Some castellanus were present in the Spanish plume above 700 mb. The small figures to the right of the sounding show the profile of wind speed in kts.

The convection over France is confined below the lid, leading to the large rise in wet-bulb potential temperature.

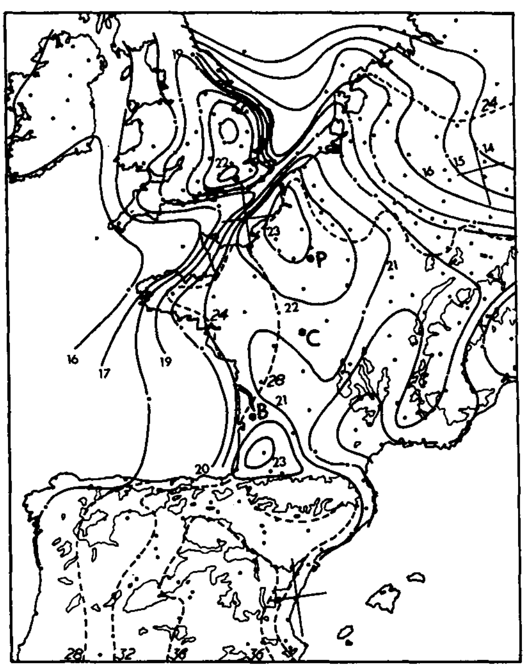


Fig. 8. Distribution of screen-level wet-bulb potential temperature (in °C), and of ordinary potential temperature (pecked lines labelled in italic figures), over western Europe, 1500 G.M.T., 5 September 1958 (see also Fig. 7). The letters *P*, *C* and *B* show the locations of the sounding stations at Paris (Trappes, 07145), Chateauroux (07354) and Bordeaux (07510). Note that the original area of maximum wet-bulb potential temperature (Fig. 7) has during the intervening 24 hours been displaced to Normandy, while another has developed over southern France.

bands lying more or less along the direction of the wind at their level; when the sources of the bands are fixed parts of the topography, their orientation in moving systems differs from the wind direction).

On the afternoon of the 4th the confinement of the small-scale convection to the lowest 150 mb underneath the Spanish air over the south and southwest of France led to the production of high values of θ_w there. The maximum value of θ_w at screen-level reached 24°C near Bordeaux (Figs. 2 and 7), and the mean in the adiabatic layer probably reached 22°C. At screen-level θ reached 29 to 32°C.

The isentropic chart for $\theta = 28$°C (Fig. 10) shows that at low levels the confluence-line was established over central France beneath the plume of Spanish air, and that the air which had been moistened over southern France had moved overnight into northern France. Accordingly, during the sunshine on the 5th a region of high θ_w developed not only once more in the south, but also over northern France, where during the afternoon the screen values reached 23°C, in contrast to the values of 20°C

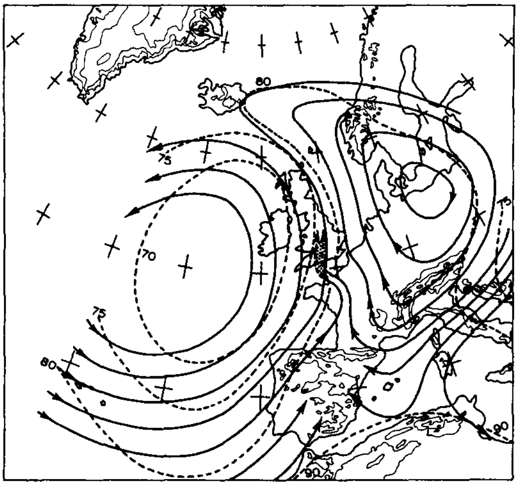


Fig. 10. Isentropic relative-flow chart for $\theta = 28°C$, 1200 G.M.T., 5 September 1958. The chart shows the flow of air from the adiabatic layer over France into the confluence zone. Over the English channel large-scale ascent leads to castellanus formation near or west of the edge of the Spanish plume, and the air leaves the isentropic surface for higher levels (streamlines become pecked, with stipple between).

or less on the previous day (Figs. 7 and 8), and the mean value in the lowest 1500 m was probably about 21°C. In this region the winds were between east and southeast, 5–10 kts, at the surface, and were also from east of south in the upper part of the adiabatic layer. The considerably backed low-level winds, consistent with a shallow trough of low pressure across central France, are characteristic of southerly flows in summer and are probably imposed by the orography and the heating pattern.

They carried the moist air into the English Channel and beyond the western edge of the plume of Spanish air aloft, where it was no longer an effective barrier to convection from below. The moist air produced fresh castellanus over mid-Channel and southern England, where a θ_w of 21°C in saturated air permitted unimpeded growth into middle levels and transformation into vigorous cumulonimbus. These moved inland and intensified into the Horsham storm, which for some hours travelled northeast with a marked squall front over which the potentially warm air still leaving the coast of N.E. France and Belgium was lifted to maintain the cumulonimbus. The storm died out soon after midnight on reaching a position off East Anglia, upwind of which θ_w at screen-level over the Continent had barely reached 20°C, and where the mean θ_w in the lowest km was therefore insufficient to sustain it.

(b) *The Wokingham storm of 9 July* 1959

Another severe storm travelled across S.E. England on 9 July 1959. It has been studied in great detail and suggested a new model for an efficiently organised cumulonimbus circulation (Browning & Ludlam, 1962; Ludlam, 1963).

Again it occurred in a cold front zone, which moved at about 10 kt from the Atlantic into the western seaboard of Europe during the 8th and 9th of July. Although the front seemed to be weak at the surface there was a pronounced trough aloft on the conventional isobaric maps, with speeds in the southwesterly flow on its eastern side reaching about 80 kts in the high troposphere, considerably more than on the occasion of the Horsham storm. Apart from the stronger wind shear the most notable difference from the earlier occasion was that the deep cyclonic vortex off Ireland was replaced by high pressure at the surface and a northwesterly flow aloft.

The flow in the upper troposphere. The relative flows and temperatures in the upper troposphere on the forward side of the trough were very similar to those on the occasion of the Horsham storm, with air from the upper part of the adiabatic layer over the western Sahara again flowing out of the desert and across Spain into a confluence in the middle troposphere over western England with a maritime stream from the Atlantic. Accordingly in the vicinity of southern England the values of θ_s in the upper troposphere were again set between 17 and 18°C by the typical properties of the Saharan air in the southerly flow.

The flow in the middle troposphere. Over central Spain on the 8th and the 9th the screen-level potential temperatures reached a little over 37°C (Fig. 11), rather higher than on the occasion of the Horsham storm, which occurred at the end of the summer season. The isentropic chart for 37°C showed that the air in the Spanish plume reached its condensation level over the English Channel; it produced castellanus over southwest England with bases at 600 mb and tops reaching 450 mb; θ_s in this layer was between 17 and 18°C.

The flow in the lower troposphere. Over

France on the 8th θ at screen-level rose to between 32 and 34°C (Fig. 11), and high values of θ_w, reaching a maximum of 23°C, occurred in the west in the characteristic pattern under the plume of air from Spain. The same pattern recurred on the afternoon of the 9th.

The isentropic chart for 31°C (Fig. 12) shows a marked distortion in the flow pattern north of Spain, which probably arose partly because of the elevated topography near the coast, stemming the flow of the shallow maritime stream into the interior, and partly by the ascent of the southerly stream of air leaving Spain with a higher potential temperature, requiring a convergence beneath. Accordingly, the air in the adiabatic layer over western France moved from the southeast beneath the Spanish plume, and over the Bay of Biscay passed under its western boundary (lying from

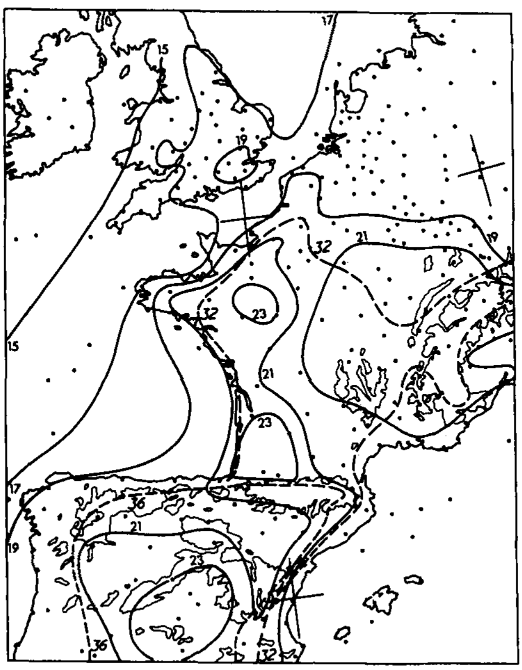


Fig. 11. Distribution of screen-level potential temperatures over western Europe at 1500 G.M.T., 8 July 1959 (wet-bulb potential temperatures: solid lines labelled at intervals of 2°C; ordinary potential temperatures: pecked lines at intervals of 4°C). The characteristic region of high wet-bulb potential temperatures over southern France is evident, and also the displaced region now over N.W. France and significant for the outbreak of the severe storm north of Brittany the following morning.

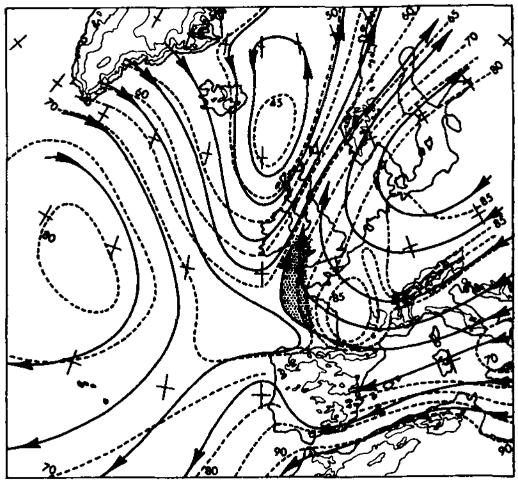


Fig. 12. Isentropic relative flow chart for $\theta = 33$°C, 1200 G.M.T., 9 July 1959. The chart shows the flow of air from the French adiabatic layer into the confluence zone, with ascent and castellanus formation in the stippled belt (arbitrarily terminated over Wales).

North Spain to the mouth of the English Channel) and rose to produce a belt of castellanus clouds which by the early morning of the 9th extended as far north as the Channel Islands and produced thunderstorms. At 0800 a thunderstorm over the Brittany peninsula intensified, and subsequently moved into and across S.E. England as the Wokingham storm. This development, as in the previous case-study, occurred close to the position of the western edge of the plume of air from the adiabatic layer over Spain, and just where the air at lower levels arrived from the region of the pronounced maximum in the screen-level θ_w over N.W. France on the previous afternoon (Fig. 11). The storm thereafter travelled over a region in which the flow near the ground was northeasterly, with the low θ_w of 16 to 18°C, but an intense squall front developed over which the cumulonimbus were maintained by the lifting of a southerly flow between about 900 and 750 mb, in which air continued to arrive from the adiabatic layer over France with a mean θ_w of 20°C.

The isentropic analyses show that the Wokingham and the Horsham storms developed in flow patterns which were essentially similar, although this was not obvious from a comparison of conventional isobaric charts.

Fig. 13. Distribution of wet-bulb potential temperature (solid lines, labelled in even degrees Centrigrade) and ordinary potential temperature (pecked lines at intervals of 4°C), 1500 G.M.T., 29 August 1961. The diagram shows (over France) abnormally high values of the latter, and low values of the former, compared, for example, with the distributions shown in Figs. 8 and 11.

(c) *The Brussels storm of 12 July 1961*

On 11th July a marked cyclogenesis occurred west of Ireland, and an associated cold front swept rapidly across Spain and western France early on the 12th. An intense thunderstorm developed towards noon in a belt of castellanus ahead of the front over N.E. France, soon reached Brussels with a marked line-squall and arch cloud, and during the afternoon moved into western Germany and decayed.

Analysis showed that this storm formed in circumstances similar to the two previously described, but, consistently with the greater westerly wind components, in a position farther east. At low levels a region of high θ_w was again present over southern France on the afternoon of the 11th, beneath a strong lid imposed by a plume from the adiabatic layer over Spain. The moist air moved northeast overnight to reach N.E. France and Belgium by the following mid-day, and then began to reach the western flank of the Spanish plume and to produce castellanus with a lower base and more vigorous growth than those already extensively present in the Spanish air. Although the upper cloudiness had prevented the morning sunshine from contributing to the low-level θ_w, the mid-day Brussels sounding shows values between 20 and 21°C in the layer below 850 mb, sufficient, as on the previous occasions studied, to sustain a severe storm.

(d) *The case of 29 August 1961*

The late summer of 1961 was notable for the absence of thundery weather over England and France. Towards the end of August a favourable large-scale situation arose, when a pronounced cold front approached, associated with a cyclone west of Ireland. During the 29th the front, and a deep associated trough on the conventional maps of the flow aloft, moved eastwards across the British Isles. Thunderstorms had seemed likely to occur near the front when it reached southeast England in the early evening, but in the event it was accompanied by only a narrow band of high altocumulus castellanus. The explanation for this behaviour, so different from that in the previous case-studies, was found in the processes affecting the lower troposphere.

In the adiabatic layer over western France (Fig. 13) the values of θ at screen-level were unusually high, reaching 36°C, and the values of θ_w rather low, lying in the range 17 to 20°C. At some places the dew-points were as low as 4 to 6°C, corresponding to relative humidities at screen-level of less than 15%. It appears that there was very little evaporation into the air over western France, and that the energy introduced into the adiabatic layer was almost entirely in the form of sensible heat. The rainfall had been unusually scanty during the previous month over the west and north of France: less than 50% of the normal (about 60 mm) over a large area and as little as 10% of normal in many places. Moreover, in the west of France the last day with a fall of as much as 1 mm had been in the first or second week of the month, and it is therefore probable that over large areas the soil had become very dry. In contrast, during the month before the Horsham storm the rainfall over practically the whole of France had been well above normal, and at many places only 2 to 3 days had elapsed since more than 1 mm fell in a day.

The soundings made at Bordeaux at midnight on the 28th and the morning of the 29th show typical Spanish air present from 850 to 700 mb with θ about 36°C. During the following afternoon θ in the adiabatic layer over western France became practically as high and the convection in this region probably extended up through the base of the Spanish plume to beyond the 700 mb level. As a result the mean mixing ratio and θ_w in the bottom km became lower in the afternoon than in the late morning, the mean θ_w falling to 18°C or less, and the condensation level of the air rising to well above 700 mb, so that no condensation was possible in the local convection or in the limited ascent accompanying the large-scale flow into the frontal zone over England. Accordingly the only clouds observed during the passage of the frontal zone across S.E. England in the evening were high altocumulus in the upper part of the Spanish plume.

(e) *Discussion*

It is well known that in western Europe the most severe summer thunderstorms occur in the vicinity of slow-moving cold fronts and in association with showering castellanus (see, e.g. Douglas & Harding, 1946). The case-studies described show that on the isentropic charts the front is clearly indicated as a confluence-line between two distinct air streams. The configuration of the flow near the front, in combination with geographical features, was peculiarly favourable for the production of intense cumulonimbus on each of the three studied occasions when severe storms developed.

Downwind of the high and arid Spanish table-land a plume of air with a high θ provided a lid to confine the small-scale convection over France to a shallow layer within which θ_w rose to values considerably above θ_s in the upper troposphere. The potentially warm air was released where it later reached the edge of the restraining plume, and the severe storms developed by the intensification of showers which propagated into the warm air supply by lifting it over a squall-front. The severity of the storms was probably increased by the efficient organisation of the cumulonimbus convection in the marked vertical wind shear found near the front. The life of the storms was restricted to several hours by the limited supply of warm air. The importance of the lid is emphasised by the absence of storms on an otherwise favourable occasion when it was made weak and ineffective by a previous drought over France.

5. Case-studies of severe storm situations over the U.S.A.

(a) *General*

Severe local storms occur frequently in the mid-western United States during the spring and early summer, when the trade wind current of the North Atlantic enters the Gulf of Mexico and moves northward across the southern states which lie east of the Rockies, carrying inland air with a high θ_w (21 to 22°C) in the lowest one or two km. At the same time θ_s in the middle troposphere over the mid-western states is as low as 16 to 18°C. The instability which leads to thunderstorms is therefore not as dependent upon recent sunshine over land as in western Europe. Nevertheless here, as elsewhere, the worst storms break out in favourable synoptic situations, especially on the forward side of troughs approaching from the west, and the greater prevalence of tornadoes in the afternoon and evening indicates that insolation plays an important part in the initiation or intensification of the cumulonimbus, if not in the provision of the instability.

The synoptic situations leading to the severe storms have been discussed by, for example, Miller (1959) and Newton (1963). On the approach of a trough from the west there is usually a strengthening of the southerly current at low levels over the Great Plains states. Soundings show the moist air to be beneath a rather strong inversion near the 800 mb level. Above the inversion θ is more than 40°C, and the lapse-rate very steep, but tall cumulus and cumulonimbus cannot form in the presence of the inversion, and a large part of the discussion in the literature is concerned with possible mechanisms for its removal. The lifting of deep columns through a pressure range of 100 to 200 mb has been suggested, and it is suspected that this occurs predominantly in the vicinity of the curious "dry-line". This is a line lying almost north-south, across which there is during the day a very rapid transition at screen-level from the dew-points of 18°C or more which

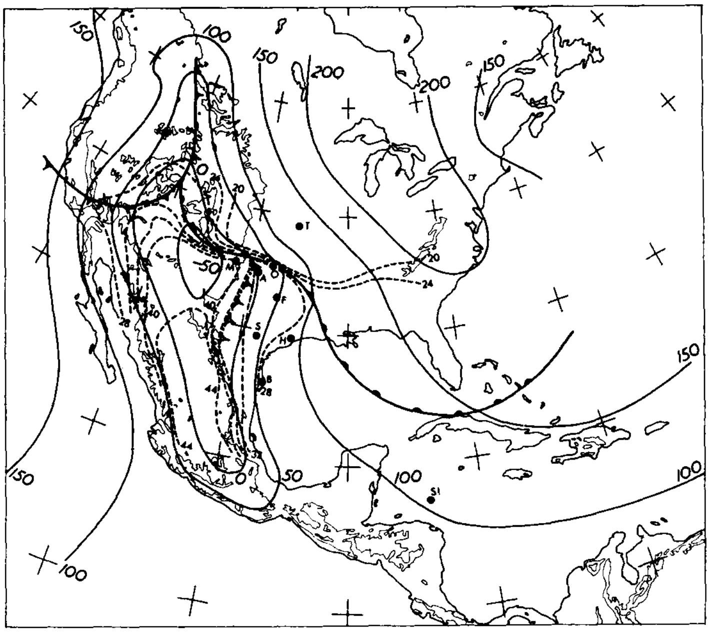


Fig. 14. Distribution of screen-level potential temperature (pecked lines labelled at intervals of 4°C) at 2330 G.M.T. on 4 May 1961 (corresponding approximately to the day's maximum values), and conventional frontal analysis with dry-line indicated as a dot-dash line with open cold-front symbols. The continuous lines are 1000 mb contours labelled in m. Letters beside full circles mark the positions of the following places, some of which are mentioned in the text; in order from north to south: *T*, Topeka; *O*, Oklahoma City; *M*, Amarillo; *A*, Altus; *F*, Fort Worth; *S*, San Antonio; *H*, Houston; *B*, Brownesville.

characterise the moist southerly flow to values of 0°C or even less on the west side. Aircraft observations in the first few thousand feet above the ground (NSSP, 1963) have shown that most of the transition occurs abruptly in an almost upright zone usually less than 5, and perhaps often less than 1 mile across. The dry-line is recognised as a boundary between the moist flow and a stream of "continental tropical" air which arrives from the southwest after flowing over the desert states. It is not regarded as a front; usually as the whole motion-system moves away eastward from the lee of the Rocky Mountains, a more important surface cold front becomes evident, introducing a colder air mass from the north, and the dry-line becomes a less prominent feature or disappears entirely. Cumulonimbus are often observed to originate near the dry-line, but a satisfactory explanation of the processes at work there has not yet been given. Case-studies have been made of two severe storm outbreaks, on 4 May 1961 and on 26 May 1962.

(b) *The occasion of the Geary storm: 4 May 1961*

The severe Geary storm, whose structure closely resembled that of the Wokingham storm (Browning & Donaldson, 1963), was one of a group of cumulonimbus which after forming over eastern Colorado late on 3 May moved across Oklahoma during the 4th.

Figs. 14 and 15 show the large-scale situation at 1730 C.S.T., according to the isobaric charts for 1000 and 500 mb. The belt of low pressure near the ground extending from the Rocky Mountains across the desert states and into Mexico is a typical springtime feature; aloft the trough was not well marked.

The flow in the high troposphere. In this region the flow in the upper troposphere is not so readily related to other parts of the general circulation as over western Europe. At levels near the tropopause θ_s is about 20°C, not much different from the mean value of θ_w in the layer near the surface which is prevalent within the tropics and available to sustain cumulonimbus convection there or to be involved in the large-

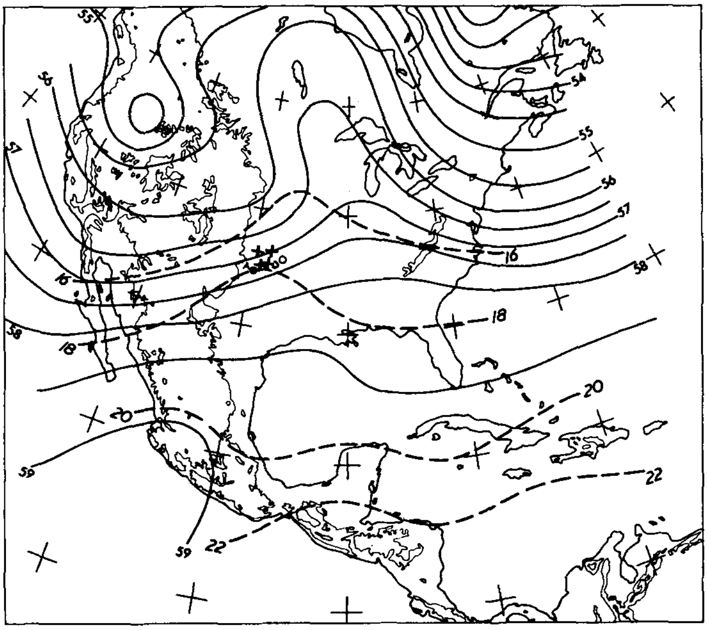


Fig. 15. 500 mb contours at 50 m intervals, labelled in hundreds of m, 2330 G.M.T., 4 May 1961. The pecked lines are isopleths of θ_s, labelled in °C. The crosses between Altus (*A*) and Oklahoma City (*O*) mark the places from which giant hail or tornadoes were reported between 1500 and 1700 hrs local time.

scale convection. The flow over the Great Plains came westwards from the Pacific without the partition into distinct streams which appears in the middle and lower troposphere.

The flow in the middle troposphere. In the middle troposphere ahead of troughs approaching western Europe the condition of the air is strongly related to the stratification imposed by the small-scale convection over the arid regions of Spain and the Sahara. Over the middle west of the United States the corresponding regions are the arid lands of the southwestern states and of Mexico.

Most of Mexico is between one and three km above sea level. On spring afternoons θ at screen level over the central area reaches about 47°C (Fig. 3). In the southwesterly winds prevailing ahead of troughs a deep adiabatic layer is found aloft downstream of this arid region, with a potential temperature of about 46°C.

The sounding in Fig. 16, made on the evening of 4 May over southern Texas, shows such a layer, with a potential temperature of 45°C. Below the base of the layer at 750 mb is a cooler but moister layer ($\theta_w = 22$°C) which entered Texas as a SSE'ly flow from the Gulf of Mexico.

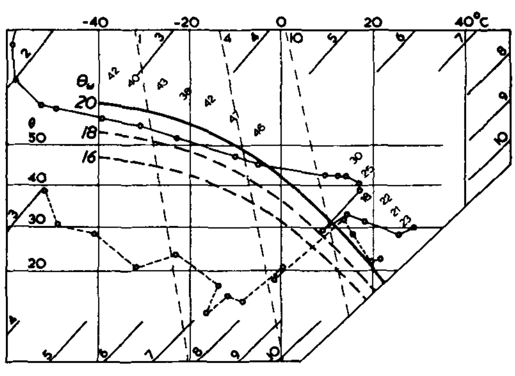


Fig. 16. Sounding for San Antonio (72253; marked *S* on Fig. 14), 2330 G.M.T., 4 May 1961. Wind speeds in kts are entered on the right of the sounding. Mexican air is present between about 750 and 550 mb.

This kind of sounding has long been recognised as a precursor of severe local storms in the American mid-west (Type 1 of Fawbush & Miller, 1954).

On the isentropic relative flow chart for $\theta = 46$°C (Fig. 17) the air stream which enters Mexico from the east turns northwards into a

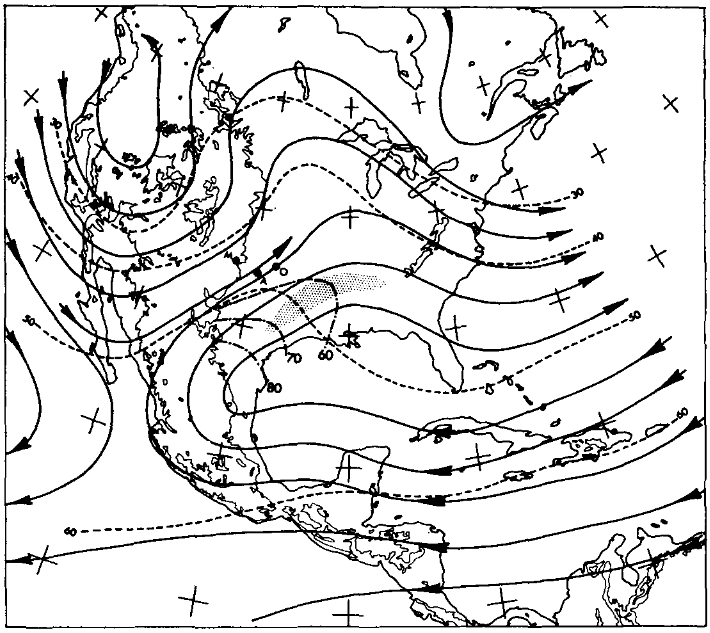


Fig. 17. Isentropic relative-flow chart for potential temperature 46°C, 4 May 1961. The pecked lines are isobars at 100 mb intervals, labelled in tens of mb. The dot-dash lines over Texas are isobars also labelled in tens of mb, which show the approximate level of the base of the Mexican plume. The stippled area shows where small amounts of altocumulus and low cirrus cloud appeared within the flow. As on Fig. 14, the letters *O*, *A* and *M* mark the positions of Oklahoma City, Altus and Amarillo respectively. The limiting streamline which denotes the edge of the Mexican plume is drawn through *O*.

confluence ahead of the trough, with a limiting streamline which passes northeastward near Oklahoma City. West of this streamline is air which has arrived from the rear of the trough. Over southern Texas the Mexican plume was about 200 mb deep, with a sharply defined base at levels between 800 and 700 mb. Nearer the limiting streamline, however, it appears to have ascended more and to be shallower, with a base between 600 and 550 mb over Oklahoma. Generally it was too dry to produce cloud during its flow over the southern United States, other than small amounts of altocumulus in a rather narrow belt originating over southern Texas (Fig. 17).

The chart for the surface $\theta = 37$°C (Fig. 18) shows the flow into the confluence zone of two major streams from the Pacific and the Caribbean, the latter appearing to skirt the highlands of Mexico, where the air is everywhere modified to a higher potential temperature, and so cannot appear on this surface. The limiting streamline of the flow from the Caribbean appears to pass just *west* of Oklahoma City and is the dry-line at about the 700 mb level. In this flow the air is rather moist, with a condensation level of about 600 mb, but west of the dry-line is the other stream, which after subsiding in the rear of the trough passed across the desert lands of the southwestern United States and northern Mexico. There a deep dry-adiabatic layer was produced with θ about 37°C and a top at about 600 mb. Since the flow from the rear of the trough sank into the layer of small-scale convection over these arid areas where practically no evaporation occurs, the air remained too dry to produce cumulus, with a mean θ_w of only about 14°C. The modified air is generally described as "continental tropical", but it reaches the confluence zone from middle latitudes.

The flow pattern in and around the confluence zone is rather similar to those previously described in the European case-studies. A striking difference, however, is that in the latter the stream which arrived from the west had been subject to small-scale convection over an ocean whose surface temperature is only 15 to 20°C.

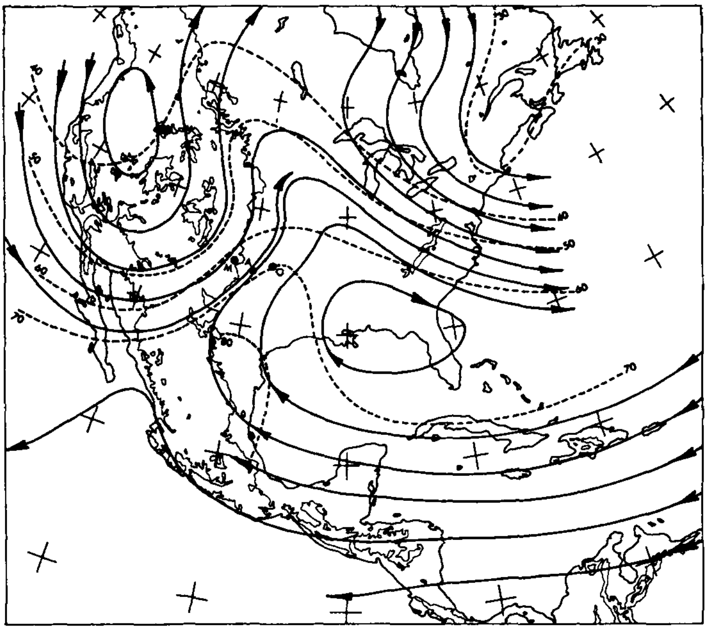


Fig. 18. Isentropic relative-flow chart for potential temperature 37°C, 2330 G.M.T., 4 May 1961. Details as in Fig. 17. The limiting streamline of continental tropical air is drawn through *A*.

This stream remained cool, in contrast to that east of the confluence line, which passed over regions where screen temperatures reached 30 to 45°C. Consequently over western Europe the confluence-line is recognised as lying in a cold front zone, and the flow on its eastern side is even at low levels subject to a large-scale baroclinic ascent which plays an important part in releasing the instability. Over the south-western United States, however, the modification of the cold air mass is so strong that the confluence-line is recognised as a dry-line rather than as a cold front. Its prevalent position is probably related to the topography of northern Mexico and the southern Rockies: the gap between these high regions may favour the ingress into the plains states of the flow from the rear of the trough, while the ascent of the air strongly modified over the Mexican highlands must require a confluence beneath the plume which it forms, lying northeastwards from near Chihuahua in northern Mexico.

The flow in the low troposphere. It might be anticipated that as the trade wind air moved inland and was subject to further heating, the small-scale convection would have intensified and extended upwards with the widespread development of showers and thunderstorms. The potential instability would then have been dissipated over a large area, probably without any violent effects.

However, inland over Texas the air moved below a lid formed by the plume of potentially warm air downwind of Mexico. The base of this plume, whose inclination is shown on Fig. 17, limited the small-scale convection to the lowest 200 or 300 mb. Accordingly as the moist air flowed inland its depth was not much increased, and the energy introduced from sunshine permitted the mean θ_w in the lowest 1 or 2 km actually to increase from its value of about 21°C at the coast to 22°C over northern Texas. On the chart for $\theta = 32$°C the flow of moist air can be followed to just north of Oklahoma City, in which vicinity it was evidently transported to the high troposphere in cumulonimbus, for farther beyond Oklahoma City the values of θ_w everywhere in the troposphere were less than 18°C, and only near the tropopause was θ_w as high as 22°C. Nearly all the severe storms occurred near the dry line, within a region of destabilization shown on Fig. 19 where the moist flow, continuing northward along a path parallel to the mountain chain of North America, reached the edge of the Mexican plume. The progressive approach to instability can be seen in soundings from near the coast, where the convection was confined below 820 mb; from central Texas where the convection reached 700 mb (and the Mexican lid was weaker), and finally from the neighbourhood of the dry-line beyond the Mexican plume (Fig. 20). Here the Mexican air was absent and the moist air intruded below the deep dry-adiabatic layer of "continental tropical" air, produced over the desert farther west; cumulus congestus formed from the moist layer, with a base between 1500 and 2000 metres above the ground, and grew through the adiabatic layer and for some distance above its top at about 550 mb, some becoming large enough to produce showers and thus to initiate cumulonimbus.

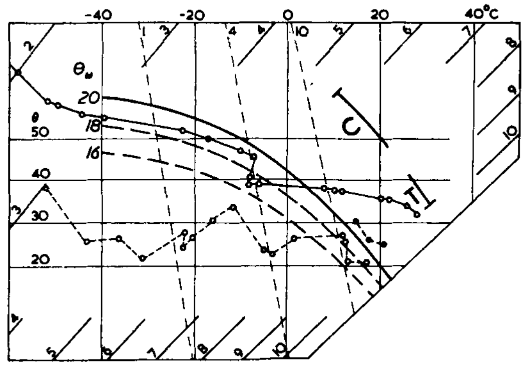


Fig. 20. Sounding for Altus (72352; marked *A* on Figs. 14, 15 and 17–19), 2330 G.M.T., 4 May 1961. The distribution of the wetbulb potential temperature in the moist trade wind layer (*T*) is shown by full circles; above is the layer of "continental tropical" air (*C*) with the characteristic potential temperature of about 38°C.

At Oklahoma City the sounding made just east of the Geary storm shows a stratification which by virtue of the steep lapse-rate and low θ_s in the middle troposphere was very favourable for the propagation of intense cumulonimbus. The adiabatic ascent of air in the lowest km (θ_w 21 to 22°C) would imply temperature excesses of as much as 7°C throughout the middle troposphere and accordingly a buoyancy greater than is normally attained in European storms, but about the same as that in the exceptional Horsham storm. The buoyancy could have been even greater in the region beyond the edge of the Mexican plume (see Fig. 20); further, the low value here of θ_w (about 16°C) in the middle troposphere, and the steep lapse-rate down to the surface, favoured the production of intense downdraughts (Kamburova & Ludlam, 1966) and therefore the *organisation* of severe cumulonimbus.

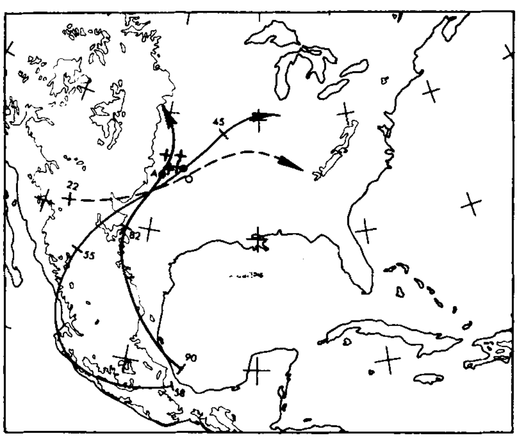


Fig. 19. Composite chart, 4 May 1961. The limiting streamlines are shown for the moist trade wind flow (thick line; height of top of layer indicated in tens of mb) and the top of the Mexican air (thin line; height marked in tens of mb). The pecked line shows the flow in the high troposphere, at about 220 mb. Notice that in the vicinity of the severe storms (crosses) the winds veer with height. The places marked by the full circles are Altus (*A*) and Oklahoma City (*O*).

(c) *The situation of 26 May 1962*

On 26 May 1962 numerous severe thunderstorms occurred near Oklahoma City and farther north-east, towards Topeka. On this occasion the surface and 500 mb flow patterns were generally similar to those on the day of the Geary storm, but the warm front was situated farther north (Fig. 21).

The flow in the middle troposphere. The relative-flow chart for the surface with $\theta = 46$°C shows that the limiting streamline of a typical plume of Mexican air passed close to Oklahoma City and south of Topeka, in which vicinity it lay just east of the dry-line at low levels (Figs. 23 and 29). The layer of small-scale convection west of the dry-line was especially warm and deep: at Amarillo (Fig. 22) the adiabatic layer extended up to 520 mb, with mean values of θ and θ_w of 42°C and 16°C respectively.

The flow in the lower troposphere. At screen-level θ was mainly between 28° and 35°C south

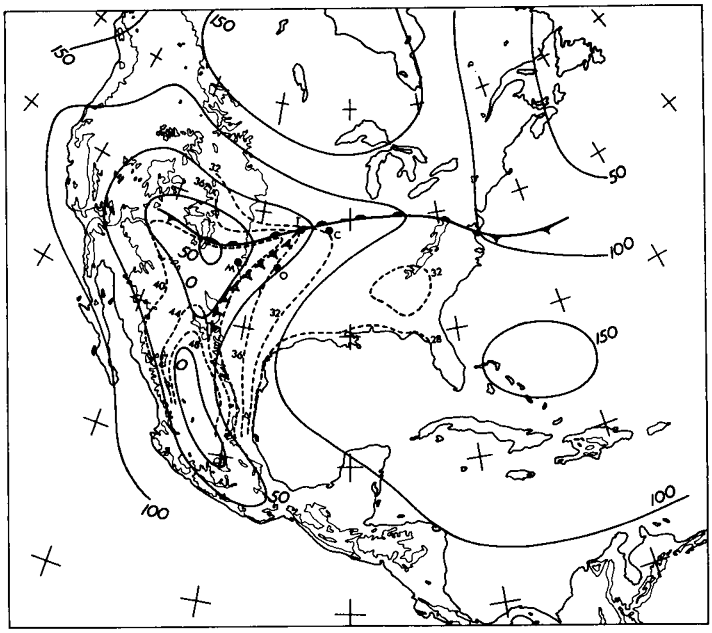


Fig. 21. Contours of the 1000 mb surface (continuous lines labelled in m), 2330 G.M.T., 26 May 1961, and conventional frontal analysis (dry-line indicated by pecked line with open cold front symbols). Also included are isopleths of screen-level potential temperature (approximately the day's maximum) at intervals of 4°C. The places marked by full circles are Amarillo (M), Oklahoma City (O), Topeka (T) and Columbia (C).

of the warm front, rising to about 36°C near the dry-line. Fig. 23 shows the relative flow in the surface for $\theta = 35°C$ (which on most soundings in the trade wind air was a little above the top of the adiabatic layer), and includes isopleths of the mean θ_w in the lowest 100 mb of the trade wind air. The diagram shows that this moist air turned northward ahead of the Mexican plateau, so that the dry-line defining its western boundary lay down-wind of the northeast corner of the plateau.

The composite flow pattern. The composite flow chart (Fig. 24) is similar to that for 4 May 1961 (Fig. 19). On each occasion the storms developed where the moist layer reached the confluence-line between the Mexican and "continental tropical" air streams. On 26 May 1962 the cumulonimbus formed during the afternoon near the dry-line and became intense in a zone up to about 100 mile wide between the dry-line and the confluence-line aloft (Fig. 29). The sounding from Topeka (Fig. 25) is similar to that made at Altus on the previous occasion (Fig. 20). As before, little rain fell south of latitude 33° N, where there was a strong lid.

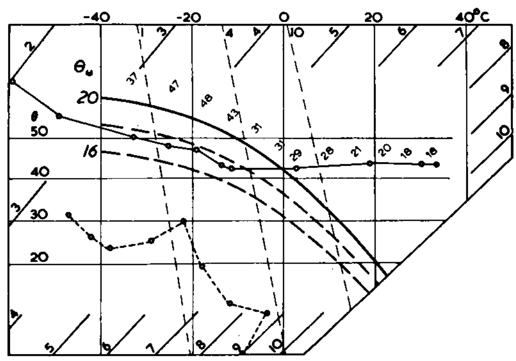


Fig. 22. Sounding for Amarillo (72363; marked M on Fig. 21) 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. An adiabatic layer of continental tropical air is present below 520 mb.

The stratification near the dry line. From our analyses we have constructed the schematic cross-section of a dry-line shown in Fig. 26. The structure suggested is consistent with previous inferences that the dry line slopes upward to the east at low levels but bends backward somewhat towards the west above about 600 mb

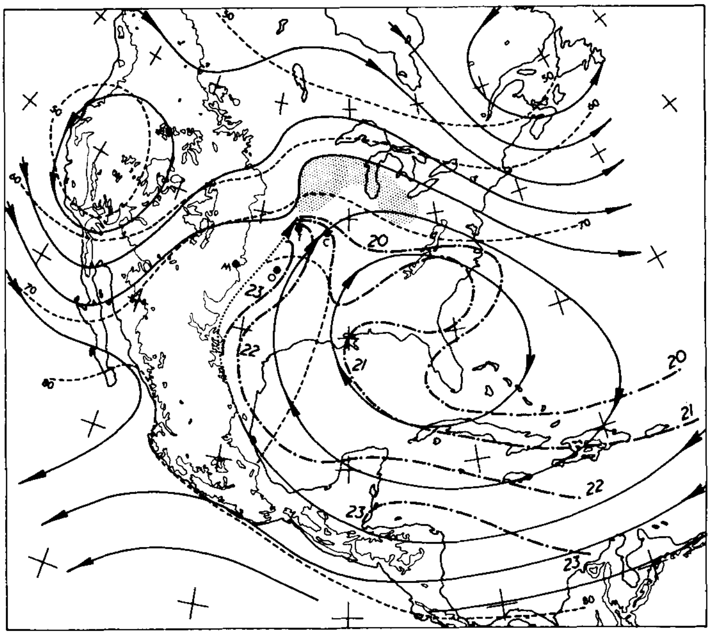
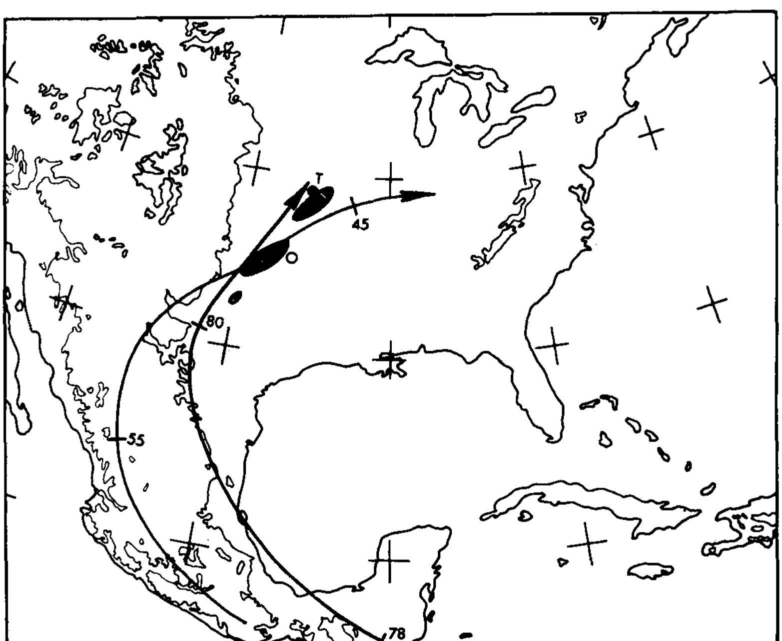


Fig. 23. Isentropic relative-flow chart for potential temperature 35°C, 2330 G.M.T., 26 May 1962. Height contours at intervals of 100 mb (pecked lines) are labelled in tens of mb. The stippled area represents a belt of middle-level cloud formed in air from the thunderstorm downdraughts after ascent over the warm front. The dot-dash lines labelled in °C show the distribution of mean wet-bulb potential in the lowest 100 mb of the tradewind flow (where this flow is not strictly in this surface, because a potential temperature a little greater than 35°C developed at the ground, it is indicated by a dotted streamline). This streamline, which delineates the edge of the moist air, passes just west of Topeka (*T*).



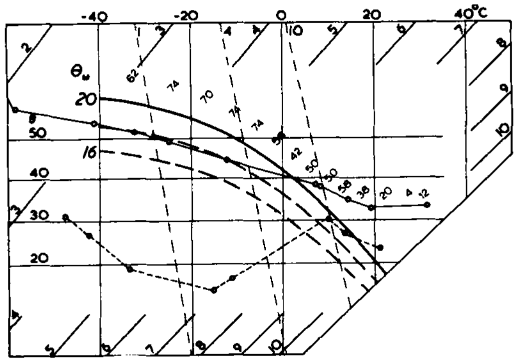


Fig. 25. Sounding for Topeka (72456; marked *T* in Figs. 23 and 24), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. The unstable stratification strongly resembles that at Altus on 4 May 1962 (Fig. 20).

(Fujita, 1958, p. 582; McGuire, 1962, p. 5). The zone of greatest instability (in respect of cumulus convection) is east of the surface position of the dry-line and near or west of the limiting streamline in the Mexican air.

In the spring, monthly precipitation decreases westward across the central plains and amounts are extremely small over western Texas and the Mexican coastal plain. The air which flows inland over the moister ground of eastern Texas receives a rather small fraction of the net radiation as sensible heat, and accordingly the mean θ in the layer of small-scale convection increases very little. On 26 May, the afternoon screen-level values of θ and θ_w well east of the dry-line were 34° and 24°C respectively; the corresponding cloud base is 1700 m (above M.S.L.), rather higher than the actual value of 1400 m estimated by photogrammetry (Fig. 27). Since the base of the lid was at about 2500 m, the cumulus here were very shallow (Fig. 28).

Farther west, however, and particularly along trajectories passing near and west of Brownesville, the duration of travel over land prior to arrival in the region of thunderstorms becomes more than one day. Because sensible heating over drier ground predominates along this part of the flow, there is a noticeable increase in θ near the dry-line; consistently, Fig. 26 shows

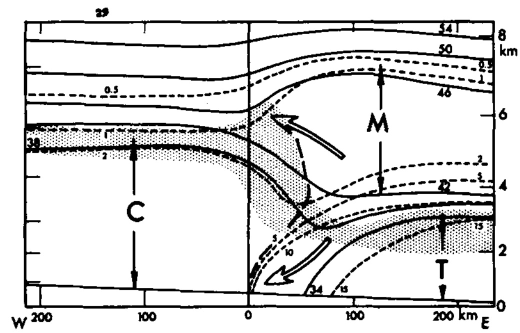


Fig. 26. Schematic cross-section of a typical dry-line, showing as a function of distance in km from its surface position the distribution of potential temperature (isopleths at 4°C intervals) and mixing ratio (pecked lines labelled 0.5, 1, 2, 5, 10 and 15 g/kg). The continental tropical air is denoted by *C*, the Mexican air by *M*, and the moist trade wind air by *T*. Arrows show the sense of the baroclinic circulations near the dry-line, and stippling shows the extent of the layer unstable with respect to cumulus convection. The heavy pecked line shows the position of the dry-line front above the ground.

that the mixing ratio decreases westward. Accordingly towards the dry-line the cumulus base rose (Fig. 27), and in a narrow zone west of Oklahoma City the clouds disappeared (Fig. 28), before re-appearing close to the dry-line

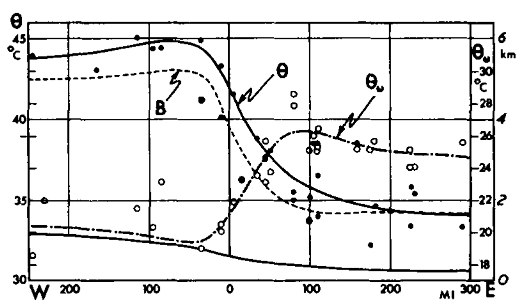


Fig. 27. Distribution of potential temperature (small circles, continuous line), wet-bulb potential temperature (large circles, dot-dash line) and adiabatic condensation level (pecked line marked *B*), of screen-level air, 2330 G.M.T., 26 May 1962. The observations are entered according to distance from the dry-line. The full circles mark the level of cumulus base estimated from photographs taken by a reconnaissance aircraft. The gradients of properties close to the dry-line may have been much stronger than indicated in the figure, based on smoothed synoptic data.

Fig. 24. Composite chart, 26 May 1962. The limiting streamlines are shown for the moist trade-wind flow (thick line; height of top of layer indicated in tens of mb) and for the top of the Mexican air (thin line; height marked in tens of mb). Also marked are the areas of severe storms (hatched areas) and the positions of Oklahoma City (*O*) and Topeka (*T*).

Fig. 28. View northwards from Oklahoma City (from position 1 on Fig. 29) at 14,000 ft, 1954 G.M.T. (1354 C.S.T.) 26 May 1962. There are scattered small cumulus over and east of the City, disappearing farther west towards the dry-line.

where the lid was not present. Ordinarily the base of cumulus is observed appreciably above the condensation level corresponding to screen temperatures, but on this occasion estimates from aerial photographs of the cumulus clusters which formed very locally near the dry-line (included in Fig. 27) indicate a base-level *lower* by about a km, supporting the view that the clusters appear only where favourable ground features induce intermediate-scale circulations in which the adiabatic layer is locally moister. The release of latent heat when some of the cumulus developed towards the congestus stage may have intensified the convection, so that after a rather long period (several hours following sunrise) in which none or only very small clouds were present, the transition to shower clouds proceeded rapidly (within an hour: see Figs. 32 and 33). The zone near the dry-line in which the Mexican air had been replaced by the "continental tropical" air was extremely unstable for cumulus convection, having an almost dry-adiabatic lapse-rate up to about 5 km.

Observations of cumulonimbus formation near the dry-line. Fig. 29 shows the positions from which the photographs reproduced in Figs. 28 and 30–33 were taken, and also the positions of the dry-line and severe storm events.

The first swelling cumulus near the dry-line formed WSW of Oklahoma City about mid-day; elsewhere along the dry-line the sky was clear. These cumulus developed into the cumulonimbus group A by 1345 C.S.T., shown in Fig. 30 at 1500, with a dome well above 40,000 ft, the height of the tropopause. A second cumulonimbus group (B: Fig. 31) developed about an hour later from the southern part of the same source region. Figs. 32 and 33 illustrate the rapid development of cumulonimbus (group E)

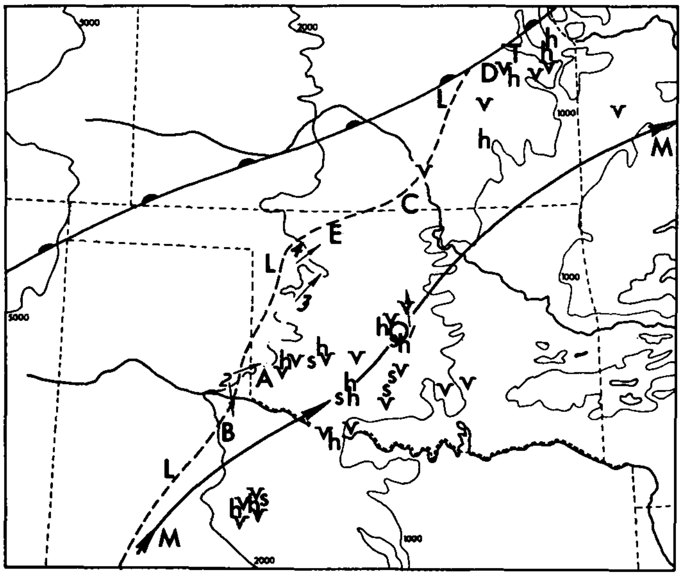


Fig. 29. Map showing position (on afternoon of 26 May 1962) of the warm front, the dry-line (*LL*), the limiting streamline (*MM*) of the Mexican air, and of Oklahoma City (*O*, in right centre) and Topeka (*T*, top right). Ground contours are drawn for heights of 1000, 2000 and 5000 ft above sea level, and the state borders and principal rivers are included. The distance between Oklahoma City and Topeka is about 280 miles. Reports of severe storm incidents between 15 and 23 hrs C.S.T., 26 May 1962, are indicated by tornado symbols and the letters *h* (large or giant hail) and *s* (destructive squalls). The italic figures *1* to *4* and the associated arrows show the positions and directions of view of the pictures in Figs. 28, 30–33. The letters *A* to *E* show the positions of cumulonimbus groups mentioned in the text and figure legends.

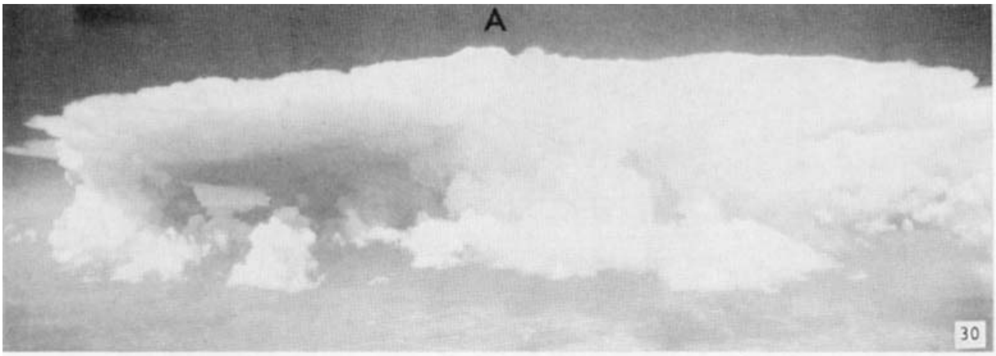
from small cumulus close to the dry-line. Similar development produced the cumulonimbus groups C and D, widely separated by spaces almost completely free of even small cumulus.

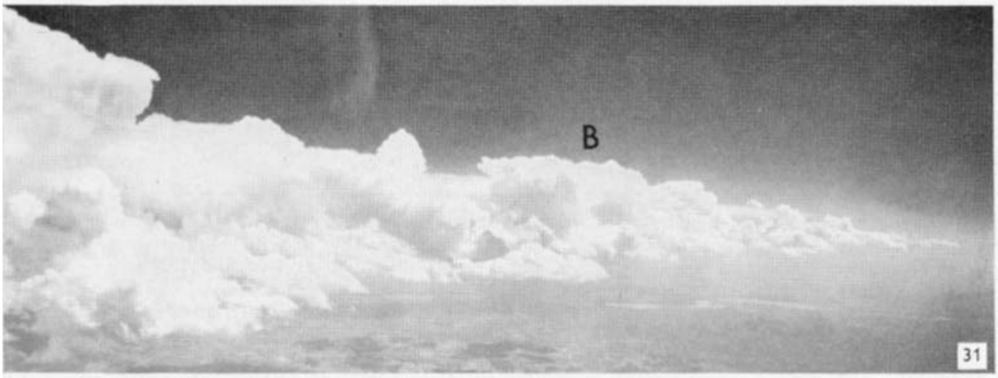
Showers developed in clouds which reached the altitude of the aircraft (400 mb), and probably produced downdraughts and squall fronts which allowed them to propagate away from their sources, to the right of the winds and so away from the dry-line and towards the region of maximum θ_w at the surface. Eventually, however, propagation may halt and the storms decay where they reach the plume of Mexican air, even though a large buoyancy could be attained by sufficient lifting of the moist air. Fig. 29 shows that on this occasion the most intense activity was associated with storm group A, which moved near but remained just outside of the region containing the restrictive Mexican lid.

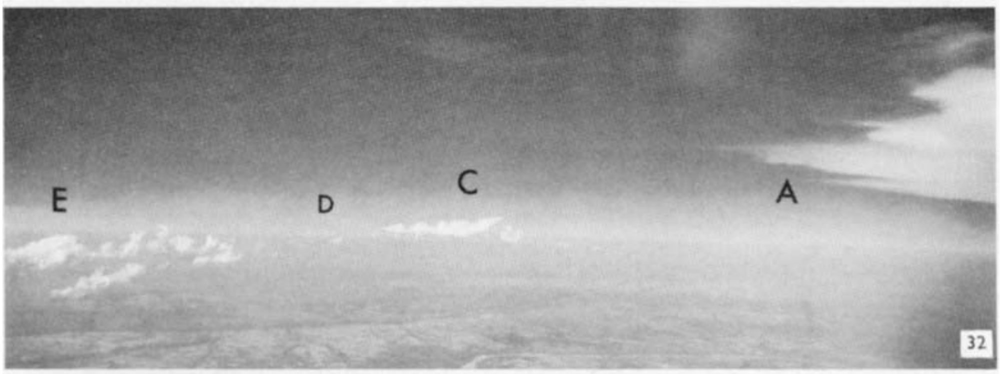
6. Conclusions

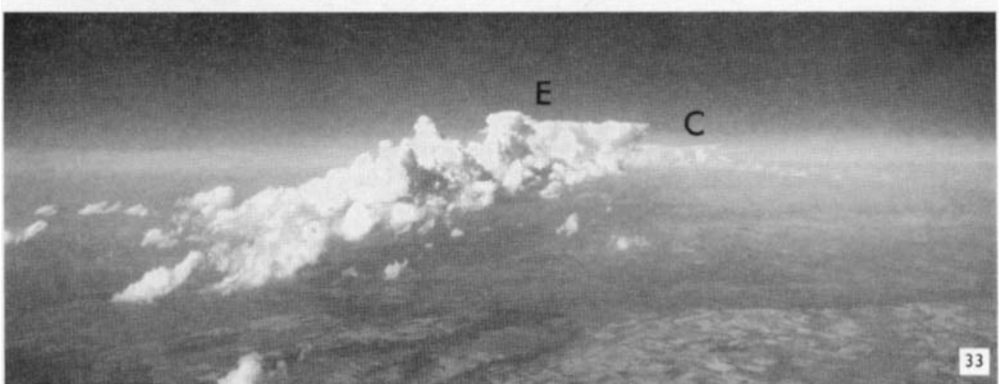
The development of intense cumulonimbus requires the presence of a wind shear which permits the attainment of a large potential buoyancy represented by a difference between θ_w at low levels (in the layer of small-scale convection) and θ_s aloft. In a given part of the large-scale flow pattern and time of year the former is the more variable, being under the immediate influence of sunshine and the kind of terrain. After one day of sunshine it may be raised sufficiently for thunderstorm development, especially where the small-scale convection may locally be confined to a shallow layer beneath a "lid" formed by the advection aloft of a layer of high potential temperature θ from a neighbouring elevated and arid region.

This favourable circumstance was an important feature in several analysed cases of severe local storms which occurred near southern England, and over the middle-western U.S.A. In the former region the storms occurred near a cold front, ahead of which an abnormally high θ_w developed at low levels over France, beneath a lid of potentially warm air from the Spanish plateau. (In one apparently favourable situation no storms developed because a preceding dry spell over France parched


A
30


B
31


E
D
C
A
32


E
C
33

the soil and led there to an unusually high θ, rendering the lid ineffective and producing an abnormally deep and *dry* layer of small-scale convection.)

In summer ahead of a trough approaching the middle-western U.S.A. a plume of potentially warm air from the Mexican plateau appears similarly to restrain small-scale convection in the moist trade-wind stream of high θ_w which persistently flows inland from the Gulf of Mexico. The western edge of this flow is marked, not as in Europe by a cold front, but by a "dry-line" which in the analyses appears as a confluence-line, west of which is a very dry current heated at low levels over desert regions after descending in the rear of the trough. Near the dry-line the relatively cool trade-wind air flows out from under the plume of Mexican air aloft and undercuts the dry air farther west, so that in a narrow zone the restraint of the lid disappears and large cumulus form locally in a stratification providing a very large potential buoyancy. Further, the dryness of the middle troposphere, and the steep lapse-rate down to the ground, favour the production of strong downdraughts, and the considerable wind shear present permits the organisation of cumulonimbus convection. The intense thunderstorms which develop after the formation of showers propagate away from the dry-line, but probably decay on entering the region where the plume of Mexican air is present as a strong "lid".

"Lids" of potentially warm air which lead to an abnormally high mean θ_w in layers of small-scale convection may be important in the production of severe thunderstorms in other parts of the world. For example, those over tropical West Africa in summer are known to be associated with an inversion aloft produced by the advection of air from the Sahara over the moist currents of the SW'ly monsoon, and the storms of the Indian monsoon develop in air in which θ_w rises significantly during flow over the Arabian Sea beneath an inversion produced by the advection aloft of a very warm adiabatic layer from Arabia (Colon, 1964, p. 194). The infrequency of severe local storms in most regions, suggesting that in general convection of lesser intensity prevents the development of extreme instability, is consistent with the view that only where this convection is impeded by a peculiar stratification imposed by geographical features can there be a storage of energy at low levels, eventually released in a violent upheaval.

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Fig. 30. View looking ENE of mature cumulonimbus (*A* of Fig. 29), from about 25,000 ft near dry-line (position 2, Fig. 29), 2100 G.M.T. (1500 C.S.T.), 26 May 1962.

Fig. 31. View looking S from 25,000 ft near dry-line (position 2, Fig. 29), 1435 C.S.T. 26 May 1962. There is intense cumulus and cumulonimbus development just east of the dry-line. The group of cumulonimbus beginning to develop anvils in the right centre of the picture is marked *B* in Fig. 29.

Fig. 32. View looking NE from 25,000 ft nearly along the dry-line, 1540 C.S.T., 26 May 1962 (position 3, Fig. 29), showing clouds of group *C* commencing anvil formation. On the original picture the tops of a similar group (*D*) can be seen clearly, although at the great distance of about 400 km. On the left congestus clouds are beginning to tower over a source-region where 30 minutes previously there were only traces of fractocumulus; 20 minutes later shower formation had begun and subsequently the cumulonimbus group *E* developed (Fig. 33). On the right is the edge of the anvil of the cumulonimbus *A*.

Fig. 33. View from 25,000 ft looking NE from position 4, Fig. 29, 1632 C.S.T., 26 May 1962. The clouds in the group *E* (Fig. 32) are beginning to produce anvils and partially obscure the more distant cumulonimbus of group *C*.

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УСЛОВИЯ СУЩЕСТВОВАНИЯ СИЛЬНЫХ МЕСТНЫЦ ШТОРМОВ

Используется метод анализа относительных потоков на изэнтропических поверхностях для изучения крупномасштабных («синоптических») ситуаций, связанных с сильными местными штормами в районе южной Англии и над Средним Западом США (включая штормы в районах Хорсхэм, Уокингхэм и Джири, которые были описаны в нескольких предыдущих работах). Штормы возникают впереди главных ложбин, вблизи от линий слияния (обычно отождествляемых с холодными фронтами над Западной Европой и с «сухими линиями» над США), где увеличение ветра с высотой способствует образованию и развитию конвекции, связанной с кучевой облачностью. Сильная неустойчивость возникает там, где граница мелкомасштабной конвекции находится не ниже, чем 1 или 2 километра (что проводит к необычно высокой влажноадиабатической потенциальной температуре), т. е. ниже слоя очень теплого воздуха, расположенного вниз по ветру от обширного засушливого плато (Испания или Мексика). Неустойчивость проявляется, когда (обратный) поток нижнего уровня в конце концов достигает ограничивающего слоя наверху. Оказывается, что существование сильных местных штормов требует особенно благоприятного сочетания географических и метеорологических условий.