

LARGE-SCALE CONDITIONS FOR THE OCCURRENCE
OF SEVERE LOCAL STORMS

by

TOBY NAHUM CARLSON

Department of Meteorology
Imperial College of Science and Technology.

A thesis submitted for the
Degree of Doctor of Philosophy in the
University of London.

June, 1965.

ABSTRACT.

The production of severe local storms depends upon the production of a substantial excess of wet-bulb potential temperature (θ_w) near the ground over the saturation wet-bulb temperature (θ_s) in the middle and upper troposphere. This is likely to develop over a moist land area when heating due to sunshine absorption at the ground is confined for a day or more to the bottom 1 or 2 km only. A principal aim of the work is to show that this happens in the presence of a stable layer acting as a thermal barrier (or 'lid') aloft, formed by an anomalously warm plume of air advected from the adiabatic layer over a hot and arid plateau. In the middle and upper troposphere θ_s in a given part of the large-scale flow pattern and time of year is relatively invariable compared with θ_w near the ground, because the distribution of the former is controlled mainly by the large-scale circulation while the distribution of the latter is influenced by the peculiarities of the terrain. The most favourable part of the flow pattern for severe thunderstorms is the southerly ahead of a trough, where the thermal barrier confining the convection may be produced while large-scale cooling by ascent permits the establishment of a low θ_s aloft.

In two investigated examples of severe local storms over southern England a pronounced rise in θ_w in the bottom 1 or 2 km over southern France is attributed to heating beneath the base of a very warm plume of air originating upwind over the Spanish table land. Over the following 24 hours the region of high θ_w was advected downstream to northern France. The instability was released in intense thunderstorms north of the Channel where the lid became ineffective. Cumulus first showered and grew to cumulonimbus near the edge of the Spanish plume and near the upwind edge of a deck of altocumulus castellanus which formed in the greater body of the French adiabatic layer. At middle and upper levels the ascending currents were found to originate in

the layer of small-scale convection over Spain and the Sahara and therefore θ_s aloft was practically identical to slowly changing values of θ_w in these regions. The front was shown to be a surface defined by limiting streamlines at the edges of the potentially warm air streams.

In another case investigated, a significant rise in θ_w near the ground failed to occur over France despite the presence of Spanish air aloft. The restraining barrier to the convection was obliterated by a rapid rise in θ due to lack of moisture in the soil following an extended dry spell.

In two investigated cases of Oklahoma thunderstorms the high θ_w near the ground was found to be provided by the northward intrusion of the trade wind moist layer ahead of the high Mexican plateau. At middle levels the easterlies entered the arid Mexican plateau and formed a deep adiabatic layer. The Mexican adiabatic layer, when advected northward, confined the convection over the plains to the bottom 2 or 3 km. This proved to be effective in conserving the high θ_w in the moist layer near the ground, while the Mexican plume ascended to reach a state of near saturation at its top; θ_s aloft thereby becoming practically equivalent to θ_w in the Mexican adiabatic layer. The edge of the Mexican and trade wind flow formed a lower tropospheric front, known as the dry line, adjoining a very warm and dry stream which was heated mainly over desert regions following its descent during flow from the rear of a major trough. Large cumulus were produced along the dry line in a region where the Mexican plume was displayed from above the moist layer. These grew to cumulonimbus which became very intense when they moved away from the dry line but remained outside the Mexican plume. The distribution of heating and the initiation of cumulonimbus along the dry line is discussed with the aid of photographs taken from an aircraft.

TABLE OF CONTENTS

	Page
ABSTRACT	3
TABLE OF CONTENTS	5
CHAPTER I : INTRODUCTION	7
1.1 Kinds of atmospheric convection	7
(1) Small-scale convection	7
(2) Intermediate-scale convection	7
(3) Large-scale convection	7
(4) Cumulonimbus convection	7
1.2 The collaboration of small- and large-scale convection	8
1.3 The stratification associated with small-scale convection	9
(1) The surface layer	10
(2) The superadiabatic layer	10
(3) The adiabatic layer	10
(4) The transition layer	12
(5) The cloud layer	12
1.4 Illustrations of the stratification accompanying small-scale convection	15
1.5 Cumulonimbus convection	17
1.6 The stratification associated with large-scale convection	20
1.7 Technique of analysis in the case-studies	22
(a) Use of isentropic charts	22
(b) Standardization of radiosonde observations	23
CHAPTER II : THE DISTRIBUTION OF DRY- AND WET-BULB POTENTIAL TEMPERATURES AT THE GROUND	26
CHAPTER III : CASE STUDIES OF SEVERE STORM SITUATIONS	37
3.1 The Horsham storm	37
(a) The flow pattern in the upper troposphere	37
(b) The flow in the lower troposphere	40
(c) Conclusions	44
3.2 The Wokingham storm	47
(a) The flow in the upper troposphere	47
(b) The flow in the middle troposphere	49
(c) The flow in the lower troposphere	49
(d) Conclusions	52
3.3 The case of 29 August 1961	53
(a) The flow in the upper troposphere	53
(b) The flow in the lower troposphere	54

3.4	Severe local storms in the U.S.A. :	
	The Geary storm	57
	(a) General	57
	(b) The occasion of the Geary storm	59
	(c) The flow in the upper troposphere	60
	(d) The flow in the middle troposphere	61
	(e) The flow in the lower troposphere	64
	(f) Hydrological aspects	67
	(g) Seasonal values of θ and θ_w	70
3.5	The case of 26 May 1962	71
	(a) The flow in the middle troposphere	71
	(b) The flow in the lower troposphere	73
	(c) The composite flow pattern	75
	(d) The influence of insolation east of the dry-line	76
	(e) Observations of cumulonimbus formation along the dry-line	81
3.6	Conclusions	84
APPENDIX 1		88
APPENDIX 2		92
APPENDIX 3		94
REFERENCES		97
FIGURES (1 - 93)		

I. INTRODUCTION

1. Kinds of atmospheric convection.

Four principal kinds of convection can be distinguished in the circulations incessantly in progress in the troposphere, which can be ranked according to their scale :

1. Small-scale convection, on scales of up to a few km, which communicates to the lower troposphere energy absorbed at the earth's surface from solar radiation. It is introduced partly as sensible heat and partly as the latent heat of evaporated water. This kind of convection is distinguished from the others in that throughout at least the lower part of the affected layer the potential density increases with height (i.e. there is static instability).
2. Intermediate-scale convection, on scales up to the continental, which consists of baroclinic circulations associated with irregular distributions of small-scale convection, imposed by the topographical features of the underlying surface. A familiar example is the sea-breeze circulation. With approach to continental scale, as in the Asiatic monsoons, it becomes impossible to distinguish this kind of convection from
3. Large-scale convection, on a horizontal scale of thousands of km, due to the general latitudinal temperature gradient, which outside the tropics effects a transport of heat not only vertically throughout the troposphere, but also from low to high latitudes.
4. Cumulonimbus convection, on horizontal scales of up to hundreds of km, which depends upon the presence of precipitation initially formed in small-scale convection. In

low latitudes it is continually at work effecting the flow of heat upward through the upper troposphere, but elsewhere it occurs only sporadically and probably contributes little to this transport.

The cumulonimbus convection might be said to play no essential part in the general circulation outside the tropics, but nevertheless it here sometimes attains an extraordinary intensity in the severe local storm, accompanied by torrential rain and giant hail, and by very strong winds at the ground. The subject of the present study is the role of the other kinds of convection in arranging for the generation of these storms.

2. The collaboration of small- and large-scale convection.

The collaboration of the small- and the large-scale convection is illustrated schematically in the vertical section of fig.1.

Here the small-scale convection is shown to occupy a rather shallow lowermost part of the troposphere into which air is led along the descending branch of a large-scale circulation. After some time the air attains a high potential temperature and enters the ascending branch, within which water vapour is condensed. Because most of it is here also precipitated the descending branch of the circulation is cloud-free ; the descent is accompanied by a radiative energy loss and a fall of potential temperature until the air again enters the region of small-scale convection.

The average magnitude of the radiative heat loss is set by the rate of absorption of solar radiation in the troposphere and at the ground, and is equivalent to a tropospheric cooling rate of between 1 and 2°C/day. Assuming that ascent in the large-scale circulations is rather rapid and nearly adiabatic, the air arrives in the upper troposphere with its potential temperature increased by the liberation of the latent heat of original water

content. If this, expressed as a mixing ratio, amounts to 10 g/kg, the increase of potential temperature is about 30°C , and since this must also be the magnitude of the decrease during the return towards the surface, the rather long descent-period of 15 days or more is indicated. The mature large-scale motion systems are observed each to occupy typically about one-fifth of the middle-latitude belt in which they occur, so that air in the upper or lower troposphere moving with the observed relative speeds crosses one system within 2 to 5 days. In the diagram, therefore, the arms of the descending circulation have been left open to indicate that the descent continues in other systems, and that on that side there is no definite boundary. It also follows that if the small-scale convection is to restore the potential temperature in such a short period, it must generally be limited to a layer occupying only a fraction of the depth of the whole troposphere.

3. The stratification associated with small-scale convection.

Each kind of convection is accompanied by characteristic distributions of the meteorological elements, such as velocity and temperature. The routine synoptic observations are made at time and space intervals adequate to allow a diagnosis of the state of the large-scale circulations and a prediction of their likely evolution.

The observations inevitably are affected by small-scale circulations and flows, and a main concern of the routine analyst is to decide what is 'representative' of the large-scale motion. The ordinary observations certainly do not define the structure of the small-scale circulations, but with the help of models developed from special observations it is possible to recognise their existence and effects.

The stratification of the atmosphere in the presence of

small-scale convection is revealed by observations made with sensitive and rapid-response thermometers which can be used on horizontal or spirally-ascending paths to determine both the average temperature and humidity distributions and also their fluctuations associated with the individual small-scale circulations.

From the results of such investigations by a number of authors (e.g. Bunker et al, 1949 ; James, 1953 ; Warner, 1963) it appears that several layers can be distinguished, characterised by certain lapse-rates of potential temperature and mixing-ratio. With the decrease in the values of these parameters from the bottom to the top of a layer indicated by the prefix Δ , these layers are as follows :

1. The surface layer, from the surface to about screen-level ;
 $\Delta \theta$ is several degrees Centigrade, perhaps more than 10, especially over arid ground in intense sunshine, and Δx is also large, often amounting to several g/kg. In this layer a considerable part of the upward heat flux may be due to radiation.
2. The superadiabatic layer, extending up to 100 to 200 m, in which the lapse-rate is appreciably greater than the dry-adiabatic and there is also a strong lapse of mixing ratio ; with reference to screen level $\Delta \theta$ is 1 to 3°C and Δx is a few g/kg, the larger values occurring over the more arid terrain.
3. The adiabatic layer, extending up a km or more to within 200 or 300 m of the cloud-base level. Within this layer θ is almost uniform, variations not reaching 1°C. In the lowest few hundred metres θ decreases upwards slightly, and probably thereafter increases. However, there is usually a noticeable and fairly steady lapse-rate of mixing ratio throughout the whole layer, and so the slight static

static stability implied for the upper part of the layer is reduced and becomes barely detectable. Further, the decrease of mixing ratio also implies that θ_w , the wet-bulb potential temperature, decreases upwards. From 100 aircraft soundings Petterssen et al (1946) found that on the average θ_w was less than the screen value by 0.7°C at 150 m, 1.0°C at 300 m and 1.1°C at 450 m. The differences may be greater over arid country. Thus over Spain, Mexico and the Sahara, the screen-level maximum θ_w , measured in settlements where the humidity may be unrepresentatively high, appears often to be as much as 6°C higher than θ_w in the upper part of the adiabatic layer.

In the lower part ascending thermals can be recognised as volumes of up to a few hundred metres across within which the virtual potential temperature (by some tenths of a degree Centigrade) and mixing ratio are higher than in the surroundings. These differences decrease with height and become inappreciable in the upper part of the layer. The upward speeds encountered, of 1 to 3 m/sec, are provided by the buoyancy due to the excess (virtual) temperature in the thermals, and in combination with them are sufficient to provide the required upward heat flux, which can at most be the net radiative flux at the surface, only a fraction of the solar constant.

The approximate uniformity of θ over most of the layer does not imply that in the individual circulations the flow is adiabatic. Rather, the evidence is that they are turbulent, containing motions on a spectrum of descending scales : there is vigorous mixing, and the stratification of θ can be regarded as a close approximation to the neutral state for a thorough stirring process. Probably within the upper part of the layer the number of thermals which reach a given level decreases as it is raised.

4. The transition layer, up to 200 or 300 m deep, may be found between the top of the adiabatic layer and the cloud-base level. It probably does not exist beneath growing clouds, into which air flows from the adiabatic layer, but is often well marked in the clear spaces between clouds, and particularly between the clusters which contain the bigger clouds. In this layer the lapse-rate of temperature decreases noticeably, so that θ may at its top be up to about 1°C higher than in the adiabatic layer, but at the same time the lapse-rate of mixing ratio increases sharply, so that the increase upwards of potential virtual temperature remains rather small. The air in the transition layer appears to have settled from the clear spaces in the layer occupied by clouds, without yet having become modified by the convection from below which, away from the cloud clusters, does not effectively extend up to the cloud-base level.

On account of the small depth of the transition layer it is not defined after a sounding has become simplified into the significant points permitted by the criteria of the radiosonde reporting code.

5. The cloud layer. Air enters the cloud bases with about the potential temperature of the adiabatic layer, but with a mixing ratio and wet-bulb potential temperature corresponding to values found at some level within the upper part of the adiabatic layer. Accordingly, the air at cloud-base has a θ_w lower than the screen-level value (by about 1.0°C in flight experience over England) and the cloud-base is about 200 m higher than the condensation level calculated from screen-level observations (e.g. on the average by 650 ft in the data of Petterssen et al (1946), varying little with the cloud-base height over the range 1500 to 6000 ft).

Within the layer occupied by cumulus clouds the lapse-rate is intermediate between the dry- and the saturated-adiabatic, so that if the cloud air rose adiabatically it would soon acquire a substantial excess temperature of some degrees Centigrade. If at the top of the cloud layer the temperature attained in saturated adiabatic ascent is ΔT higher than that given by dry-adiabatic ascent, the actual temperature is found to be higher by about $2\Delta T/3$.

Although the updraught speeds encountered in cumulus a km or more deep are in the range 3 to 6 m/sec, about double those found in thermals within the adiabatic layer, from the reasonable requirement that the upward heat flux should be no more than that near the ground it can be anticipated that the temperature within a cloud does not exceed the clear air temperature at the same level by more than fractions of a degree Centigrade. This seems to be supported by flight observations, which are hampered by difficulties due to wetting and evaporation at the thermometers, but which indicate that excess temperatures of more than 1°C are rare. Clearly, therefore, the cloudy convection is also strongly subject to mixing; the importance of the heat of condensation is shown by the increase of θ with height, and the importance of the mixing by the decrease with height of θ_w within the clouds, or of θ_s everywhere, where θ_s , the saturation potential temperature, is defined as the wet-bulb potential temperature corresponding to a state of saturation at the observed temperature.

In the cloud layer the mixing ratio and θ_w also decrease with height. Over the ocean, especially in the absence of showers, the general level of the cumulus tops is often well marked by a distinct change in the lapse-rates of θ , θ_s and x . The mixing ratio decreases rapidly upon entry into the layer unaffected by recent small-scale convection, in which the air has for some time been settling from the upper troposphere, and θ_s begins to increase with

height, the lapse-rate becoming less than the saturated adiabatic and therefore clearly incompatible with the presence of cumulus convection.

Over land the irregular terrain and the consequent intermediate scale circulations produce a more complicated structure in the convection layer. For example, the cumulus tops reach levels considerably higher over hills than over the plains, and the modified air may be brought by the winds aloft across the plains, where the soundings are usually made. This makes the soundings more difficult to interpret, especially when they are not represented in detail and are not accompanied by accurate cloud observations. Extreme examples are found down-wind of extensive mountains, where the adiabatic and cloud layers characteristic of the high ground may be found superimposed upon those due to the local small-scale convection.

Although the near-uniformity of θ over most of the convective layer below the cloud-base level represents an approximately neutral state for a vigorous stirring process, the neutral distribution of the mixing ratio is approached only when the convection continues for a sufficient length of time in a confined layer. The upper part of the layer may then become more generally saturated and the cloud a complete cover. Extensive sheets of stratocumulus are produced in this way over the oceans in winter anticyclones, when the upward extension of the convection is hindered or prevented by high values of θ aloft associated with strong subsidence. Radiation from such low-level cloud tops then plays an important part in determining the stratification; the strong radiative heat sink there probably becomes important for the persistence of the convection and in producing an extremely abrupt increase of θ and decrease of x and θ_w just above the cloud. Over land it is unusual to produce a complete cloud cover, and the strong static stability aloft which is required to confine

the convection to a shallow layer is more commonly provided by the advection above it of a layer of high θ from higher ground upwind.

4. Illustrations of the stratification accompanying small-scale convection.

Fig.2 shows a sounding made about noon on a summer day inland over southern England. The measurements form one of a series by Miss Woodward in a towed glider, which on the gliding descents traversed a number of thermals. The sounding shows the characteristic strong lapses of θ and x in the lowest 50 to 100 m and the very nearly neutral stability in the adiabatic layer. The bases of the small cumulus were found at 1315 to 1340 m, and the transition layer has its base nearly 300 m lower. In a cloud-base and in lower thermals θ was about 0.2°C higher than in the adiabatic layer and about 1.0°C less than at screen-level, while x was about 6.9 g/kg compared with values down to 5.8 g/kg in the adiabatic layer and 8.2 g/kg at screen-level. In the air at cloud-base and in the thermals θ_w was 12.4°C , about 1.6°C less than at screen-level, and the observed cloud-base was just 200 m higher than the adiabatic condensation height of the screen-level air.

Fig.3, from Malkus (1954), is an aircraft sounding made in a spiral ascending path through a trade-wind current, a little distance inland over the island of Puerto Rico. The adiabatic and cloud layers are easily recognisable, but the distributions of temperature and mixing-ratio with height are rather irregular, especially in the upper part of the cloud layer. The irregularities may be attributed to the inhomogeneities produced by the ascent of individual clouds, which are drawn out by wind shear into thin quasi-horizontal sheets. They may be noticeable for some time after the evaporation of the clouds, because there is little diffusion or disturbance in the statically stable clear air, unless there is a fresh cloud penetration, but their positions and magnitudes are unlikely to be reproduced on soundings separated by hours or tens of km.

Fig.4, also from Malkus (1962), is of another sounding in the western Atlantic trade-winds, this time over the open sea.

Fig.5 contains radiosonde and aircraft observations made about noon on a summer day near Flagstaff, Arizona (Cunningham, Glass and Carlson, 1964). The radiosonde sounding made at the airport is compared with flight observations made in and around the cumulus clouds clustered some 15 miles to the north over mountains rising up to 5,000 ft above the airport level. Although the stratification is only broadly indicated by the conventional simplification of the sounding to a few significant points, the characteristic layering is evident. The top of the layer affected by cumulus convection, in this case probably during the previous day upwind, is marked by a small inversion at about 510 mb, where earlier a shallow broken layer of altocumulus was observed. Such clouds commonly occur during the morning and evening, and are probably formed when a damp layer composed of spread-out cumulus tops is lifted in an orographic wave or an intermediate-scale circulation. They usually disperse when fresh convection clouds reach or surpass their level and induce a slight subsidence in their surroundings. Above the altocumulus level θ_s increases with height and there is a sharp decrease of mixing ratio. The cumulus encountered by the aircraft over the mountain mostly had tops below about 550 mb, and the diagram shows that on horizontal traverses through them only small temperature changes were found. Outside the clouds the mixing ratio is considerably higher than indicated on the radiosonde sounding at the same levels. This may be regarded as due to moistening by the dissolution of previous clouds, or perhaps more probably as evidence of the intermediate-scale circulation and anabatic flow over the mountains, which, combined with the small-scale convection, there intensifies the upward flux of water vapour. In consequence of the intermediate-scale circulation θ_w near the ground may be as much as 2°C higher over hills than over the plains, so that there even the cloud-base θ_w may be higher than the screen-level θ_w over the plains - by about 1°C in this example, and by between 1 and 2°C in another

studied by Ludlam (1962), where the hills were only about 800 ft higher than the low ground. The convection over the mountain was intensifying, and the aircraft investigated some isolated taller clouds in which precipitation was encountered. In the rising tops somewhat larger temperature excesses were measured, and in these cloud tops the rise of air from the cloud base must have been more nearly adiabatic than in the smaller clouds. The decrease in the value of θ_w or θ_s compared with that at the cloud-base amounted to little more than 2°C over a distance of 3 to 4 km rather than about the same value in only 1 km.

5. Cumulonimbus convection.

This last observation is consistent with much evidence that following the development of precipitation the convection becomes more vigorous and efficient, with at least a proportion of the cloud air ascending more nearly adiabatically. The rapid penetration of cumulonimbus towers into the high troposphere soon after shower formation has been remarked upon by many observers, for example, Thorkelsson (1946) in Iceland, Craddock (1949) in Malaya, and Howell (1960), also in the tropics. From the air the distinction often seems obvious between the cumulus convection occupying a hazy layer with a well defined top in the lower troposphere, and whose numerous peak towers continually penetrate a few hundred metres higher, and the cumulonimbus convection represented by a few massive clouds whose summits reach or surpass the tropopause before subsiding and spreading out into extensive anvil clouds.

It seems that the invigoration of the convection is related to the production of a cold downdraught by the partial evaporation of the precipitation outside or beneath the clouds. In the presence of the chilled air the flux of air upward into the cloud becomes part of a baroclinic overturning in a deep layer of the lower troposphere, and in this larger-scale flow the effectiveness of mixing between the ascending air and its surroundings is reduced.

The cumulonimbus develops as a succession of cloud towers which penetrate into the high troposphere, before spreading into a rather persistent frozen anvil cloud. The top of the layer of air spreading out is sharply defined by the upper almost level surface of the anvil, but its base is obscured by precipitation, which especially near the parent cloud conceals the true thickness of the layer, probably little more than a km.

It is interesting to consider the equilibrium which evidently exists between the air in the outflow of the anvil and its surroundings, for given only the anvil height and the general temperature at this level it provides an indication of the θ_w with which the cloud air arrives in the high troposphere, not otherwise easily obtained. The comparison of this value with that inferred for the cloud base shows how closely the ascent approximates to an adiabatic process.

For the corresponding problem with cumulus convection it is unnecessary to consider the precise reference process, since the only alternatives are the true and pseudo-adiabatic reference processes for condensation into liquid water, while for vertical displacements of a few km the two processes lead to final temperatures which differ by less than 0.5°C (Saunders, 1957). In cumulonimbus, however, conditions are much more complicated. Cloud water is converted into precipitation whose trajectories differ from the air trajectories, so that in some parts of the cloud the condensed water concentration is less, and in others more, than that according to any simple reference process. Moreover, freezing and condensation into ice, as well as upon liquid water, occur in variable proportion and position. The extreme reference processes are again the simple (liquid) water pseudo-adiabatic of the aerological diagram, and the ice-saturation true adiabatic, which leads to temperatures in the high troposphere greater than indicated in the former process by 2 to 5°C (as θ_w defined by the former increases from 10 to 20°C ;

see Saunders, 1957, p.346). Probably a large part of the cloud water is removed as precipitation before entering the upper part of the cloud or freezing, so that the attainment of these higher temperatures is not to be anticipated. Moreover, if a concentration of as much as a few g/kg (as suggested by the observations quoted by Jones, 1960, p.192) of small ice particles remain in the anvil outflow, equilibrium will be reached with a cloud temperature higher than that in the surroundings by about 1°C . Accordingly an ascent practically without mixing is indicated on the aerological diagram by a temperature at the level of the anvil top which is within 1 or 2°C of that on the pseudo-adiabatic through the point representing the cloud base, that is, by a θ_s not more than about $\frac{1}{2}^{\circ}\text{C}$ below the θ_w (or θ_s) at the cloud base. In Appendix 1 several examples are given of occasions when measurements of cumulus and cumulonimbus heights are available, which show that this criterion is approximately fulfilled even when the cumulonimbus are not especially intense.

During nearly adiabatic ascent a large buoyancy, corresponding to a temperature exceeding that of the surroundings by at least one and perhaps several degrees Centigrade, must be acquired near the top of the cumulus layer. Even when reduced by the weight of a nearly adiabatic concentration of cloud water, this may be an order of magnitude greater than the buoyancy characteristic of the cumulus, and consequently the updraught speeds in cumulonimbus are substantially greater than in cumulus.

When there is a favourable wind shear in the vertical the cumulonimbus becomes especially well organised and persistent. The flow of the updraught may be so nearly steady that the rise and fall of individual cloud towers can no longer be recognised in the summits, and the anvils become very large, spreading for hundreds of km in the direction of the upper winds. The smooth ascent of air is not significantly impeded by mixing processes, so that it may attain speeds of up to 50 m/sec and penetrate several km into the stratosphere before settling back into the

anvil (Ludlam, 1963).

In tropical regions, where cumulonimbus are the main agency of heat transport through the upper troposphere, the typical strati-fication is nearly neutral for cumulus convection in the lower troposphere, and over the whole troposphere is nearly neutral for cumulonimbus convection. The main seats of the convection are inland hills and islands, and accordingly the mean value of θ_s in the high troposphere is close to the mean value of θ_w near the ground in these places during the afternoon.

6. The stratification associated with large-scale convection.

Outside the tropics the state of the upper troposphere is governed by the large-scale 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 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 through the upper troposphere, and the large-scale convection can be regarded as having an efficiency which makes the cumulonimbus convection unnecessary.

In the summer the principal heat sources in the northern hemisphere become the land masses somewhat north of the tropics; monsoon circulations develop over Asia, North America and North Africa, and the cumulonimbus convection and its own nearly-neutral stratification advance into higher latitudes. The large-scale convection persists, however, presumably mainly because the great thermal inertia of the oceans maintains a considerable latitudinal

temperature gradient, and north of about the 40th parallel the mean state remains unfavourable for the penetration of cumulonimbus into the high troposphere (figs. 6, 7).

Since the large-scale convection plays a strong part in modifying the stratification near and north of this parallel, it also firmly controls the location and intensity of the cumulonimbus convection. This control seems to be exercised not so much by variation in the potential temperature (θ_s) in the upper troposphere, as by the provision of favourable wind shears for the organisation of the cumulonimbus and, in combination with geographical peculiarities, the production of abnormally high values of θ_w in the lowermost layers. The modification of the mean stratification is illustrated by the meridional cross-sections of figs. 8, 9, representing respectively the flow behind and in advance of a trough approaching the west coasts of Europe and North Africa. In both of these sections the distributions of θ_s with height in middle latitudes do not differ appreciably from that in the mean (fig.7). In the rear of the trough, in the equatorward flow over the ocean, the low-level θ_w is dependent essentially only upon the sea surface temperature and is therefore everywhere less than that found in low latitudes and, except in the lowest km (occupied by the small-scale convection), less than the θ_s above. In advance of the trough the higher values of temperature and θ_w produced overland in sunshine permit the possibility of cumulonimbus convection. In particular, 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 over NW France of intense cumulonimbus which travelled into southern England as the Wokingham storm, the subject of a case-study to be described later.

7. Technique of analysis in the case-studies.

a) Use of isentropic charts.

In the analysis of the events leading to the production and release of the instability required for cumulonimbus convection it is desirable to introduce some simplifications without losing sight of the effects of surface heating and evaporation on the state and motion of the air.

According to the previous discussion of the large-scale motion system, represented schematically in fig.1, over most of it 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 thermodynamic histories 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 . 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 outside the layer of small-scale convection to justify the construction of trajectories over periods of up to 2 or 3 days. This proves a great advantage if it can be assumed that over such a period the large-scale motion system is in a steady state. Since 3 days is rather short compared with the entire life of the system, this will usually be a reasonable assumption where no rapid development or decay seems to be involved. However, its justification is really to be found in the usefulness of the analyses made with its help. In practice, the main advantages are found to be

1. 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 ;
2. the complication of examining such a sequence of maps is

avoided, by compressing the essential features into one map. The constraint of ensuring a reasonable time evolution which is normally used in preparing a sequence of maps is not altogether relinquished, for data obtained before and after the principal observation time are entered upon the steady-state map with an appropriate displacement and are used in making the analysis.

In some parts of the steady-state isentropic map the assumptions are clearly incorrect, but not on this account misleading. In particular, where the trajectories lie within the layer of small-scale convection, and have only some statistical meaning, potential temperature and mixing-ratio can be seen to increase along them. However, the general flow and the significant processes are usefully indicated, the observations even providing a measure of the non-adiabatic effects. Further, where the air enters the rather narrow region of large-scale ascent and condensation occurs, it leaves the surface, and may be replaced by air from beneath, so that the true trajectories are in some doubt. On the whole, however, the analysis on isentropic surfaces with the steady-state assumption is much more illuminating and more readily grasped than conventional analyses, providing a more obvious indication of the large-scale convection and of the effects of the small-scale convection. We shall use this technique almost exclusively in the case-studies of severe storms.

b) Standardisation of radiosonde observations.

The intermediate scale circulations have effects upon the radiosonde observations which are most marked and usually recognisable in the lower troposphere. The observations at high levels from a large area are found to have inconsistencies which cannot be attributed to such circulations, but only to errors of measurement or communication. Errors of measurement may be due to departure of the radiosonde sensors from their calibrations because of launching jolts or other reasons, but probably a more serious

source of error is neglect or incorrect assessment of the effect of solar radiation on the thermometer. This kind of error produces a systematic difference between day-time and night-time soundings by which its average value can be assessed, but because it probably varies considerably according to cloud cover and albedo it is not possible to remove the error from an individual sounding.

Other inconsistencies appear in comparing soundings from different countries, presumably because of the use of various types of radiosonde and handling or correction procedure. In general the errors tend to produce an inaccuracy in the reported geopotential height of the isobaric surfaces which increases with height. The most practical way of reducing them seems to be the method recommended by Hawson and Caton (1961).

In this method the errors are assessed at the 100 mb level, which is the highest level at which there are likely to be abundant wind observations. On the isobaric chart the average geopotential height reported by the several stations in the compact British network is arbitrarily chosen as the correct value in the centre of the network. A set of contours is then drawn to fit this value and to provide geostrophic winds which as nearly as possible agree with the reported winds. This procedure is justified by the observation that, at least in summer over much of the middle latitude area involved, the 100 mb wind speed and acceleration are small. It is remarkable that the constructed contours agree rather well with the reported geopotential height at other British (and American) sounding stations abroad (such as Malta, Gibraltar, Nicosia and Tobruk, in the Mediterranean). On the other hand, discrepancies will be apparent at other stations, which can be used to estimate a mean virtual temperature error between the surface and the 100 mb level. This is distributed over the layer according to the likely kind of error ; in the analyses which follow a geopotential error of x hundreds of m at 100 mb involved a correction to the temperatures reported on the sounding of $x^{\circ}\text{C}$ at 600 mb, $2x^{\circ}\text{C}$ at 400 mb and $2.5x^{\circ}\text{C}$ at 250 mb. The value of x was commonly 1 and occasionally as much as 4.

All the soundings were adjusted in this way before the data were extracted from them and entered on the isentropic charts. This procedure certainly seemed to improve the analysis.

Unfortunately nothing much can be done to overcome the unresponsiveness of humidity sensors, which is especially great in the low temperatures of the high troposphere. Generally cloud observations and some tendency for the reported humidity to increase with height, even though the actual value remained low, have been used to infer the presence of middle- and high-level layers in which the air is actually nearly saturated, with corresponding values of θ_w . The deduction often seems to be confirmed by the presence in the layers of minor wind speed maxima which indicate the greater acceleration of the potentially warmer streams of air.

II. THE DISTRIBUTION OF DRY- AND WET-BULB POTENTIAL TEMPERATURE AT THE GROUND

The buoyancy which may develop within cumulonimbus is proportional to the difference between θ_s in the middle and upper troposphere, and the (higher) θ_w present near the ground.

The buoyancy which is important can be judged to be that representative of a flow not much less than a km across, such as will usually be fed by an adiabatic layer several hundred metres or more thick. Apart from the very shallow superadiabatic layer at the ground, it is this layer which contains the potentially warmest air, and the value of θ_w significant for cumulonimbus convection can be anticipated to be the average value of θ_w within it, and not the somewhat higher value obtained from afternoon screen-level readings.

The value of θ_s aloft is determined by the characteristics of the large-scale circulation, and in a given part of a motion system probably varies little from an average seasonal value. Severe storms are associated not with an abnormally low θ_s aloft, but with an abnormally high θ_w near the ground. It is interesting to consider how θ_w in the low troposphere is influenced not only by the large-scale circulation pattern, but by the energy introduced by the small-scale convection and its partitioning between sensible and latent heat.

In North America the severe storms occur in areas where the low-level flow is part of a persistent monsoonal circulation carrying inland air which for several days has been subject to small-scale convection over the southern part of the North Atlantic and the Caribbean Sea. This air enters the southern United States during the early summer with the rather high mean θ_w of about 22°C in the lowest km. The conditions in this part of the world will be considered later, and for the present the discussion will continue considering the circumstances over and near western Europe. Here, in contrast, the ocean and sea surfaces are relatively cool

and the small-scale convection which leads to high values of θ_w occurs over land during sunshine. From experience it seems that in western Europe thunderstorms are rather closely associated with the rise of low-level θ_w produced by one day's sunshine, breaking out in the afternoon or the succeeding night hours and lasting up to several hours. This diurnal course is typical in the tropics, and so also in these middle latitudes the outbreak may be repeated after the next day's sunshine, if the large-scale circumstances are still favourable. Accordingly, the increase of θ_w which occurs in the layer of small-scale convection, and in particular in the adiabatic layer, as a result of one or sometimes two days' sunshine is important in explaining the occurrence or severity of the thunderstorms.

In order to estimate this increase it is necessary to know the amount of energy introduced into the convection layer. Its energy balance during the day is expressed in terms of the energy balance at the surface, where practically all the available measurements have been made, by the equations

$$E = S + H + L - R_L \quad (1)$$

and

$$H + L = R_n - G, \quad (2)$$

where

- E is the energy introduced,
- S is absorption of short wave radiation, a term which is probably small and henceforth neglected,
- H is sensible heat transferred from the ground,
- L is the latent heat of water vapour introduced by evaporation at the ground,
- R_L is the net loss of heat by long-wave radiation, which depends upon the moisture and temperature stratification and will be given its typical

average value corresponding to a local cooling rate of $1^{\circ}\text{C}/\text{day}$ (Möller, 1951) throughout the layer.

R_n is the net radiation at the ground, for which some measurements are available, and

G is the heat transferred into the ground.

Lettau (1951) first showed that G and H are related by the conductive capacities of the soil and of the lower atmosphere :

$$G/H = \rho_s c_s \sqrt{\kappa} / \rho_p c_p \sqrt{K_H} , \quad (3)$$

where ρ_s and ρ are respectively the densities of the soil and the atmosphere, c_s and c_p the corresponding specific heats, and κ and K_H are corresponding (molecular and eddy) diffusivities of heat. Priestley (1959, p.107) and Dyer (1961) show that G/H is usually a small fraction and implies an appropriate value of K_H to be between about 10^5 and $10^6 \text{ cm}^2 \text{ sec}^{-1}$. Conductive capacities of various kinds of surface material are given in the table :

Table 1
Conductive capacities of typical surface
materials

Material	Conductive capacity cal, cgs units $\times 10^2$
Dry sand	1.2
Dry soil	1.4
Heath	3.3
Clayland	3.6
Dry ploughed soil	4.2
Marshy	5.8
Air (convection layer)	$10 (K = 10^5) \text{ to } 10^2 (K = 10^6 \text{ cm}^2 \text{ sec}^{-1})$
Air (by night)	0.1
Water, still	4
Water, strong current	10^3

The values in the table show that fertile and wet materials absorb considerably more heat than dry soil and so tend to reduce the diurnal temperature range. Given the type of soil G can be expressed in terms of H , and the problem is reduced to one of estimating R_n .

Monteith (1962) has shown that the short wave radiation R_s incident at the ground during the day in clear weather over Britain is given by the expression

$$R_s = a - b (1 - 50) \quad \text{ly} \quad (4)$$

where l is the latitude in degrees and a and b vary with time of year as shown in the table :

Table 2
Constants of Equation (4)
after Monteith (1962)

<u>Month</u>	<u>a</u>	<u>b</u>
May	686	5.8
June	731	2.1
July	692	1.7
August	585	6.8
September	425	9.1

Values of R_n can be obtained from R_s by another empirical formula, provided by Monteith and Szeicz (1961) :

$$R_n = R_s (1 - A) / (1 + \beta) + L_o \quad (5)$$

where the surface albedo A and the parameters β and L_o are dependent mainly upon the soil type, as shown in the table :

Table 3
Values of parameters in Equation (5)
after Monteith and Szeicz (1961)

<u>Type of soil</u>	<u>A</u>	<u>~~β~~</u>	<u>L_o</u> ly
Bare, dry	0.17	0.39	- 140
Plains	0.25	0.25	- 110
Grass	0.25	0.22	- 140
Crops	0.15	0.18	- 160

Although both expressions were developed for conditions in Britain they appear to be reasonably valid for any middle-latitude region. They imply that on cloudless mid-summer days in middle-latitudes about 350 calories are introduced into the lower troposphere over each cm^2 , producing a rise of θ_w which is inversely proportional to the depth of the convective layer and which amounts to about 2.5°C if this depth is 150 mb, considering the long-wave loss R_L which is equivalent to a fall of θ_w by about 0.5°C in twenty-four hours.

That the rise of θ_w is directly proportional to the net introduced energy, however the insolation is partitioned between E and H , is evident if it is expressed as an increase of equivalent potential temperature and readily confirmed on a thermodynamic diagram. The table below shows how the increment of θ_w due to an income of 350 ly in one day is related to the depth of the layer throughout which it is distributed, assuming that the long-wave radiative loss is equivalent to a reduction of $\frac{1}{2}^\circ\text{C}$ in θ_w :

Table 4
Influence of depth of convective layer upon increase
in θ_w due to income of 350 ly in one day

Depth of convective layer	Increase in θ_w
<u>mb</u>	<u>°C</u>
100	4.0
150	2.5
200	1.8
250	1.3
300	1.0
400	0.6
500	0.4

If all the energy were introduced in the form of sensible heat the rise of mean potential temperature θ in the same layer would amount to about 9°C, whereas if it were wholly in the form of the latent heat of water vapour there would be no rise in θ . Since the convection layer extends up to some level where the potential temperature θ (or its equivalent should condensation occur) has the same value as that produced near the ground the actual increase in θ_w depends rather sensitively upon the partitioning of the introduced energy between sensible and latent heat. Over an arid soil the depth of the convection layer is likely to become large and the average increase of θ_w small, whereas over a water surface or moist ground the smaller rise of θ will tend to confine the convection to a shallower layer within which there will therefore be a larger rise in θ_w .

Experience in estimating and measuring evaporation shows that over vegetated soils H is rather small compared with E . Penman, (1956, p.18) and Fruitt and Aston (1963, p.74) (see also Appendix 2), for example, speak of values as low as 0.1 in agricultural situations, while the energy (expressed in ly/day) corresponding to mid-summer daily amounts of evaporation have about the same magnitude as the

net radiation at the ground :

Table 5
Evaporation measurements
(mainly from Penman, 1956)

<u>Location</u>	<u>Period</u>	<u>E</u> <u>ly/day</u>	<u>Method of</u> <u>measurement</u>
Tennessee, U.S.A.	July (15 years)	210	Water balance
Munich pine forest	June-July (10 days)	390	Energy balance
Harz mountains	July (6 years)	190	Water balance
Lake Hefner	August	325	Tank
Eastern England	May	120 to 220	Flux observations (Pasquill, 1949)

While the vegetation is able to draw upon moisture present at some depth below the surface the daytime rate of evaporation continues to be high, but after some time without rainfall the soil becomes parched, the vegetation wilts, and the evaporation becomes greatly reduced. Dyer (1961) has published some interesting measurements of the ratio $L/(L + H)$ made throughout a summer season over pasture land bearing coarse grass in clumps, in Australia. These showed that after a rainy day $L/(L + H)$ quickly rose to values between 0.65 and 0.75 and declined to values about 0.5 during several subsequent days without appreciable rain. Similarly, $L/(L + H)$ decreased sharply with declining values of soil moisture content in the days following periods of rainfall at O'Neill, Nebraska, during Project Prairie Grass (Halstead, 1954, p.358). This soil was considerably more arid than is typical of well-cultivated or good pasture land in western Europe. Even here, however, the evaporation rate may fall sharply after three or four weeks without summer rain, towards the low values

typical throughout the summer in arid regions subject to intense insolation and little rainfall, such as most of North Africa, northern Mexico, and large parts of the Meseta of Spain. Consequently these regions become notable for unusually high values of θ in the convection layer, corresponding to high daytime surface temperatures and deep dry adiabatic layers. These features are illustrated in figs. 10 and 11, which show the distributions of mean afternoon screen-level potential temperature during summer months over the United States and Mexico and over west Europe and North Africa.

Fig. 11 is based on long-term July averages over Europe and North Africa. It includes also a mean streamline over North Africa representing the flow in the layer between 850 and 700 mb, which shows that there was a persistent flow from the Mediterranean into the Central and Western Sahara. Although the flow pattern varies somewhat with height in the lower troposphere and pronounced intermediate scale circulations associated with the mountains make it difficult to infer trajectories, it seems reasonable to compare the stratifications at Tamanrasset, in the Central Sahara, and at Tobruk, on the Mediterranean coast, and to attribute the difference to the effect of several days of sunshine over the desert. The mean soundings (fig. 12) suggest that little moisture was added to the atmosphere, and that the modification was primarily by the introduction of sensible heat at a mean rate of about 200 ly/day during 5 days. At the end of this period the adiabatic layer has a potential temperature of about 43°C and extends upwards to above the 600 mb level. When the adiabatic layer has attained such a great depth the radiative heat loss becomes a large fraction of the energy gain by short-wave absorption, and the sensible heat introduced into the layer produces only a small rise in the mean potential temperature. Accordingly, the potential temperature within the adiabatic layer and at screen-level becomes rather uniform over a large area throughout the Central and Western

Sahara, and it becomes possible to speak of a characteristic Saharan air mass, which at this time of year has a θ of about 43°C and a θ_w which is just over 18°C near the surface and 17°C near its top, between the 600 and 500 mb levels.

Inspection of fig.11 shows that the afternoon screen-level θ , and therefore the potential temperature of the adiabatic layer, reaches rather uniform values over smaller regions according to the latitude and the characteristic topography. For example, over the Meseta of Central and Western Spain, a hot, dusty stretch of sparsely vegetated table-land at a mean elevation of nearly a kilometre, the screen-level θ is generally between 35 and 37°C . On the other hand, over the more fertile and lower ground of France the corresponding values are much lower, between about 26 and 28°C . The high values of θ over Spain are not due to prolonged insolation, since the usual wind speeds ensure that air does not remain over the country for more than one or two days, but are typically associated with the high ground elevation. It appears that the air which is heated over extensive high ground is not generally that which approaches at a low elevation, but that from some higher level which has a greater initial potential temperature. Thus the flows which enter Spain at about the level of the table-land have already a potential temperature of over 30°C , and the increase of temperature which occurs during the daytime is exaggerated by the general aridity of the soil.

An important consequence of this phenomenon is that plumes of air of high potential temperature, which according to the extent of their sources may be broad and deep, are found downwind of arid and elevated regions. Where these plumes lie over fertile, low-level regions they effectively restrict the local small-scale convection to the layer beneath them. It frequently happens, for example, that in a southerly airstream the convection over southwest France is confined to the lowest 1 to 2 km below a plume of Spanish air. An example of the stratification which arises is given in

fig.19, showing a sounding made at Chateauroux at mid-day. Air from the Spanish adiabatic layer, with a characteristic potential temperature of about 36°C and the rather low θ_w of less than 17°C , is observed above 800 mb, restricting the local small-scale convection to the lowest $1\frac{1}{2}$ km and permitting the θ_w in this rather shallow layer to rise above 19°C . The Spanish air will be said to impose a 'lid' upon the convection over France, in consequence of which the θ_w in the lowermost adiabatic layer attained values over 20°C later on this day (cf. also fig.9). This value was significantly higher than the mid-troposphere values of θ_s , which were about or less than 18°C , and therefore favourable for cumulonimbus development if some mechanism of releasing the deep convection could be found. This situation, and the severe storm which occurred, is discussed in more detail later.

In general the greater the rise of potential temperature at the lid the more effective it will be in confining the local convection to the lowermost layer. The thermals may be expected to penetrate some distance into the uppermost layer and by mixing before subsidence to incorporate some of it into the lower layer. In this way the depth of the adiabatic layer and the potential temperature in it may increase beyond the values to be anticipated by a simple modification of an initial sounding corresponding to the introduction of the sensible heat H . Ball (1960) has shown that the additional heating depends primarily upon H and may approach it in magnitude. In the cases of prime concern to us, where the convection is confined to a rather shallow layer over fertile ground by an intense lid, H itself is rather small and the effect probably not significant. Over arid regions where H is large the local cloud convection is unimportant: not only does the great depth of the convection layer reduce the θ_w which governs the cloud convection, but the latent heat of condensation becomes inefficiently available owing to the great height of the condensation level, or condensation may not occur even at the top of the layer. Over the Sahara, for example, the

cumulus may have a base above 600 mb even over the mountains, while elsewhere there is usually no cloud at all even though the convection layer reaches to near the 500 mb level.

Over the seas surrounding Europe and North Africa the mean potential temperatures are low, emphasising the predominance of the small-scale convection over land. The relative coolness of the sea during the summer ensures that on the average no sensible heat is provided to the atmosphere, and in the virtual absence of small-scale convection there is very little evaporation. The complicated surface pattern of moisture and potential temperature distribution associated with the variable geography proves useful in the analysis of the air flow, and imposes a detailed stratification in the vertical.

III. CASE-STUDIES OF SEVERE STORM SITUATIONS.

1. The Horsham storm.

This storm was one of the several most severe to occur over Britain during this century. Some aspects of it have been described by Ludlam and Macklin (1959, 1960). The storm formed in the English Channel during the afternoon of 5 September 1958, moved inland and intensified, producing a tornado and hailstones of tennis ball size near the Sussex town of Horsham. The velocity of the storm was about $235^{\circ} 15$ kt, increasing to 25 kt during the most intense phase ; it passed south of London into the North Sea, there dying out at about midnight.

In the first days of September a moderately deep warm sector depression had approached the British Isles from the Atlantic and had become almost stationary 300-400 miles off southwest Ireland. The central pressure was about 984 mb on the 2nd, but thereafter the depression occluded and slowly filled ; the cold occlusion entered England from the southwest early on the 3rd, accompanied by a narrow rain belt with local thunderstorms. On the 4th the rain belt became quasi-stationary across Central England, and the front extended into eastern and southern France and Spain, but it became poorly defined and by the morning of the 5th had disappeared from the official weather analyses.

a) The flow in the upper troposphere.

The isentropic analyses have been made on the assumption that the flow pattern was practically stationary from the 3rd to the 5th of September, so that the charts contain the reported winds and the streamlines or trajectories have on this occasion about their correct position relative to the ground topography.

On the chart for $\theta = 52^{\circ}\text{C}$ (fig.13) the position of the front

over western Europe is clearly recognisable as a zone of rather rapid height change extending through northwest Spain into England, this zone corresponding to a concentration of isotherms on a conventional isobaric chart. The front is also characteristically seen (Appendix 3) 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 and to the south of the deep cyclone west of Ireland. The exact position of the trajectory which defines the edge of the stream from the Sahara is not certain because southwest of Spain and off the African coast it passes through a region of sparse data, and unfortunately at this time the most useful soundings at Lisbon and Madeira usually seemed inconsistent with the others available. However, it is clear that this trajectory must be drawn into the region somewhat south of the intertropical front over the southern Sahara.

The intertropical front over west Africa separates the moist southwesterlies derived from the S.E. trade winds of the southern Atlantic and the dry easterlies which enter the Sahara from the eastern Mediterranean. In the month of September the front reaches its most northerly position, extending from the west coast in the neighbourhood of Dakar approximately between the 16th and 18th parallels, and reaching the southern slopes of the Ahaggar and Tibesti mountains (see, for example, Hamilton and Archbold, 1945). The moist flow has a mean θ_w of 22 to 23°C, and where, more than about 200 miles south of the surface position of the front, its depth reaches several thousand feet, travelling line squalls are common and the stratification in the troposphere is that associated with persistent cumulus and cumulonimbus convection. At Niamey (13½°N, 2°E), which at this season is within a few hundred miles of the surface front, the wind above about 850 mb is easterly, and at 700 mb is frequently from somewhat north of east. At about this level the air is evidently often derived from the adiabatic layer over the Sahara ; sometimes, between about 700 and 600 mb, the

characteristic dry-adiabatic lapse-rate with a θ of about 43°C is found, but commonly, as on the sounding for 4 September 1958 (fig.14), the lapse-rate appears to have been modified by condensation. The condensation within the Saharan air probably occurs following some ascent over the front, and produces altocumulus clouds which develop into castellanus. The prevalence of such clouds in the layer from about 600 to 500 mb, with a potential temperature of about 43°C at their bases, has been described by Dhonneur (1957). The corresponding θ_w in the layer reaches values of 18° to 19°C , which are rather higher than those of 17 to 18°C characteristic of the central Sahara, and suggest that the air has previously risen from near the surface in the intermediate scale circulations over the Tibesti mountain range, which lies like a barrier across the northeasterly flow entering the Sahara. It is interesting that in the western Sahara the flow near the surface becomes westerly just north of the intertropical front and that here too, therefore, it leaves the region only after ascending over the mountains. At Tamanrasset, which lies in the lee of the Ahaggar mountains, the soundings often show the air at the top of the adiabatic layer to have a rather high mixing ratio and θ_w , corresponding to representative values near the surface and clearly indicating the importance of the convection over the mountain (see, for example, fig.25). Accordingly, where the middle-latitude circulation into Africa reaches its most southerly position, somewhat south of the surface position of the intertropical front, the air is found in middle levels with a θ_w of up to 19°C .

On the occasion under discussion the Niamey sounding shows the small-scale convection in the Saharan air to have extended up to the surface for $\theta = 52^{\circ}\text{C}$, at 460 mb ; above this level the air becomes drier and more stable. On the isentropic chart the soundings at Niamey and Fort Trinquet show that where the flow turns northward near the coast of West Africa the air has a condensation level which is probably somewhat above the 400 mb level. This is

consistent with the appearance of a belt of cirrus over northwest Spain, where the air has risen to between about 380 and 350 mb. The cirrus extends northward into eastern England, with a rather well defined western edge corresponding to the position of the limiting trajectory. (A separate area of extensive cirrus over Ireland and within the cold air mass is probably produced by lifting of air previously moistened by cumulonimbus convection over the sea southwest of Ireland.) In the high troposphere in the vicinity of the front over England the value of θ_s is limited to the 19°C or somewhat less (considering the effect of some radiative heat loss) set by the value of θ_w in the flow from the south.

On the isentropic charts for lower surfaces the limiting trajectories for the southerly flow appear to enter West Africa in positions successively farther north, and on the chart for $\theta = 43^{\circ}\text{C}$ (fig.15) the air leaving the uppermost part of the adiabatic layer over the Sahara has the significantly lower θ_w of about 17°C . Its condensation level is between 480 and 450 mb, and is barely attained over southeast England. Consequently this is also about the value of θ_s found there at these levels.

b) The flow in the lower troposphere.

On surfaces of potential temperature lower than 43°C the front is the confluence of the westerly maritime air flow and a north-easterly flow which enters the Mediterranean from S.E. Europe before turning into an easterly and then a southerly stream. Over land near the confluence small-scale convection modifies the streams and complicates their structure.

Fig.16 shows the flow in the surface for $\theta = 33^{\circ}\text{C}$. On this surface the maritime stream, which has been moistened by small-scale convection over the ocean, flows across Portugal and northwest Spain without substantial modification, but over the eastern half of Spain the ground potential temperatures on the afternoons of the 4th

(fig.17) and 5th (fig.18) reach about 35°C , 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 from northeast Spain marking the base of the layer modified by convection over eastern Spain, and shrinks in width as the layer rises and produces a belt of extensive castellanus clouds with bases at about the 700 mb level. The sounding at Chateauroux (fig.19), on the right of the belt, has a layer with a practically dry-adiabatic lapse-rate and θ of 34°C from 850 to nearly 700 mb, above which level the steep lapse-rate and high humidity suggest that the castellanus extend upward to at least 600 mb and perhaps locally to above 500 mb. At Bordeaux (figs. 20, 21) the base of the Spanish air appears to be between 850 and 800 mb and on the mid-day sounding the castellanus, with bases probably somewhat below 700 mb, extend up to about 550 mb. Although neither sounding indicates a state of saturation near 700 mb, it generally seems in these situations that the clouds occur in moist plumes related to intermediate-scale circulations in the convection layer. These plumes are rarely evident on the soundings, either because the humidity element is insufficiently responsive or because the soundings do not happen to traverse them. (The castellanus clouds are well known to appear in bands lying more or less along the direction of the wind at their level. Our analysis indicates that since the sources of the plumes are fixed parts of the topography the orientation of the bands will differ from the wind direction in moving flow-systems.) On the other hand, lapse-rates considerably different from the dry-adiabatic are a strong indication of modification by cloudy convection ; unfortunately the conventions of the reporting codes and the necessity to limit the number of significant points on the sounding will often prevent a confident analysis.

The soundings at Chateauroux and Bordeaux indicate that at the castellanus bases θ_w reaches up to 19°C , nearly the values observed at screen-level near the high ground in eastern Spain,

and sufficient to permit a rather vigorous cloud growth which led to some showers reaching the ground in the castellanus belt. A slight thunderstorm occurred in mid-afternoon at Tours, northwest of Chateauroux, but rainfall stations in the castellanus belt reported falls of only a mm or less. More generally within the layer of Spanish air θ_w was about 17°C . In southeast England it is less easily recognisable on the sounding at Crawley (fig.22), but seems to have a base at 770 mb, and from there to 530 mb the lapse-rate is somewhat steeper than the saturated-adiabatic and the air is rather moist, with θ_w varying between 16 and 17°C .

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 (fig.17), and the value in the upper part of the adiabatic layer probably reached 22°C . At screen-level θ reached 29 to 32°C .

The isentropic chart for $\theta = 28^{\circ}\text{C}$ (fig.23) shows that at low levels the confluence was established over central France beneath the plume of Spanish air, and indicates that overnight the air which has been moistened over southern France had moved 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. Here the screen-level values rose to 23°C during the afternoon, in contrast to the values of 20°C or less on the previous day (figs.17, 18). Unfortunately no sounding was made at Paris, well within the moist region, but the reports of cumulus with bases at 3000-5000 ft between Paris and the Normandy coast indicate values of θ_w at cloud base of about 21°C and mean values of between 21 and 22°C in the lowest 1000-1500 m. In this region the winds were between east and southeast, 5-10 kts, at the surface, and also from east of south in the upper part of the adiabatic layer. This configuration of the low-level winds, corresponding to a shallow trough of low pressure across central France, is characteristic of

broadly southerly flows and is probably imposed by the orography and the heating pattern. Over northern France the convection was confined beneath the plume of Spanish air, but where the moistened layer flowed out into the English Channel from the Normandy coast it began to emerge from under the western edge of this plume. In the same region pronounced large-scale vertical motion, which was greatest near the limiting streamlines in the southerly flows, raised and cooled the base of the Spanish plume, so that, as seen on the sounding for Crawley in S.E. England (fig.22), it was less hindrance to the ascent of air from below. On this sounding a θ_w of 21°C in saturated air at 5,000 ft would permit the unimpeded growth of cumulus into the middle levels, and of cumulonimbus to the tropopause with great buoyancy throughout deep layers in which θ_s was between 17 and 19°C . According to the drawing on fig.24 the moist air rose and condensed into extensive castellanus over mid-Channel and southern England.

It is not clear exactly how the condensation was produced in air which arrives over the Channel and southern England from northern France. It is possible that the air was lifted over downdraught fronts associated with slight showers from castellanus in the plume of air from Spain, but more probably the larger-scale ascent in this strongly baroclinic zone was sufficient to produce the condensation. It is significant that the showers developed near or just beyond the limiting streamline of the Spanish plume, where this air was no longer able to suppress the growth of clouds formed in the air beneath. During the afternoon castellanus with bases at 8000 ft and showers were observed over southern England as far west as 3°W , and an aircraft off the coast at 3°W reported a shower from castellanus with a base at 7000 ft and tops at about 15,000 ft. Castellanus with base at 8000 ft were also observed in the Channel Islands and slight showers occurred at Cherbourg. Near the Isle of Wight, more directly downstream of the θ_w maximum over France east of the Greenwich meridian, activity was more intense. After 1000 hrs sferics were located over the Channel

and the Normandy coast, and in the early afternoon thunderstorms occurred over the Isle of Wight and the neighbouring south coast of England. After 1530 hrs associated radar echoes which appeared over the east of the Isle of Wight moved northeastward into England and rapidly intensified. Probably when the showers had become more intense lifting over their squall fronts became more vigorous, and with the organisation of the cumulonimbus convection in the wind shear the storms were able to propagate northeastwards into the region in which the Spanish air was still present. The severe storm which developed moved into an inland area where the screen-level Θ_w had risen to 22°C , and the extreme intensification was probably related to this more favourable low-level value, which was several degrees higher than that over the sea. After leaving this area, south of London, the storm decreased in severity, but for some hours it continued to travel towards the northeast, somewhat to the right of the middle- and upper-level winds, 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 the screen-level maximum Θ_w over the Continent had barely reached 20°C , and where the mean Θ_w in the lowest km was therefore probably insufficient to sustain it.

c) Conclusions.

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 with bases at 6,000 to 10,000 ft (see, e.g., Douglas and Harding 1946). On this occasion the front had been omitted from the official analyses because of the difficulty of locating it. On the isentropic analyses for surfaces well above the ground it is clearly indicated,

less as a narrow belt of strong temperature gradient than as a line of confluence between two distinct air streams. The configuration of the flow near the front, in combination with the topographical features, is peculiarly favourable for the production of severe cumulonimbus.

Downwind of the rather high and arid Spanish table-land a plume of air with a high θ confines the small-scale convection over France to a shallow layer within which θ_w rises to values considerably above the θ_s characteristic of the upper and middle troposphere. The potentially warm air is released over southern England where its flow leads it beyond the edge of the restraining plume, and the severe storm develops by the intensification of a shower which propagates into the warm air supply by lifting it over a squall-front. The severity of the storm is probably increased by the efficient organisation of the cumulonimbus convection in the marked vertical wind shear found in the vicinity of the front. The life of the storm is restricted to several hours by the limited supply of potentially warm air.

According to the analyses the values of θ_s in the upper troposphere near cold fronts over England and France are determined by the values of θ_w characteristic of the warm air flows from the south, just ahead of the confluence line. In these streams the air in the upper troposphere arrives from the convection layer over the southern Sahara, while that in the middle troposphere arrives after modification over Spain. The following typical values of θ_s suppose that the streams arrive in a state of near saturation, and that loss of heat by radiation during one or two days' travel reduces θ_w by about $\frac{1}{2}^{\circ}\text{C}$ compared with the values found in the source regions. The values are based upon soundings at Saragossa in northeast Spain during 1962, and at Miami, Fort Trinquet and Aoulef in 1958.

Table 6

Probable θ_s aloft in southerly flows over
England and France ($^{\circ}\text{C}$)

	<u>May to</u> <u>June</u>	<u>June to</u> <u>July</u>	<u>July to</u> <u>August</u>	<u>August to</u> <u>September</u>
High troposphere (air from southern Sahara)	18	18	18½	18
Upper troposphere (air from Sahara)	16	17	16½	16
Middle troposphere (air from Spain)	14	17	18	17

2. The Wokingham storm.

Another major hailstorm travelled northeastwards across S.E. England on 9 July 1959. It was not quite as severe as the Horsham storm and was not accompanied by tornadoes, but it produced hailstones of golf-ball size near the Berkshire town of Wokingham and was at the time of writing the worst storm to have occurred over England since the Horsham storm. Its internal structure has been studied in great detail and provided a new model for an efficiently organised cumulonimbus circulation (Ludlam, 1963).

Again on this occasion the storm occurred in a cold front zone which had moved from the Atlantic into the western seaboard of Europe during the 8th and the 9th of July. The front seemed to be weak at the surface, and the associated cyclone over northern Norway was shallow (central pressure about 1007 mb), but a trough with pressures somewhat below 1015 mb extended along or just ahead of the front southwestward to southern Spain, and there was a pronounced trough aloft on the conventional isobaric maps with wind 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. Above the first few thousand feet the flow aloft over southern England was southwesterly rather than southerly, but the most notable difference between the two occasions was that the deep cyclonic vortex off Ireland on the earlier was replaced on the latter by high pressure at the surface and a northwesterly flow aloft.

During the 8th the eastward speed of the trough was about 12 kts north of 50°N , diminishing southwards to about 6 kt at 34°N , and these velocities were subtracted from the observed winds during the preparation of the relative-flow isentropic charts.

a) The flow in the upper troposphere.

The flow in the upper troposphere on the forward side of the trough was very similar to that on the occasion of the Horsham storm.

On the chart for $\theta = 43^{\circ}\text{C}$ (fig.24) again air from the upper part of the adiabatic layer over the western Sahara flows out of the desert and across Spain into a confluence with a maritime stream from the Atlantic. The streamline which marks the edge of the flow from the desert passes over the mouth of the English Channel at about the 500 mb level, and condensation farther north produces altocumulus over western England which were observed to have this base-level by reconnaissance aircraft. In this flow θ_w , and hence θ_s at 500 mb over southern England, had about the typical value of between 17 and 18°C . The nearly adiabatic lapse-rate between 680 and 560 mb on the sounding at Bordeaux (fig.30), can be regarded as evidence of the presence of the Saharan air.

Again a moister flow from the southern Sahara, with streamlines passing between Niamey and the surface position of the intertropical front, seems responsible for a belt of thick cirriform cloud which formed west of Spain and extended into western England and the Irish Sea, and thence into Scandinavia. The base of this cloud was observed by reconnaissance aircraft to be at 450-400 mb southwest of Spain and 300 mb over western England. The position of the western edge of this cloud belt, which is a limiting streamline on a higher isentropic surface, is shown on fig.24. The value of θ_w within the cloud over England was a little over 18°C , and would probably have been about 19°C over the southern Sahara. It is interesting that on this occasion there is clear evidence of a convergence towards the Ahaggar and Tibesti mountains in the surface flow over the Sahara, and the soundings at Tamanrasset, in the lee of the Ahaggar, on both the 7th and the 9th (fig.25) indicate condensation, and a θ_w of about 19°C , at the top of the adiabatic layer. This evidently is the effect of the intermediate scale-circulation over the mountains, for in the middle of the adiabatic layer the mixing ratio is less and θ_w has its typical value of about 18°C . On the 9th practically all stations between the intertropical front and the twelfth parallel reported large amounts of altocumulus.

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.

b) 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 (figs.26, 27), about 3°C higher than on the occasion of the Horsham storm, which was at the end of the summer season. On the 8th screen-level values of θ_w reached 23°C and cumulonimbus and thunderstorms formed over some of the mountains in the afternoon. The sounding at Zaragoza at midnight (fig.28) shows that extensive cumulus convection has reached 650 mb, and the lapse-rate above again becomes steeper than the saturated adiabatic up to 450 mb.

At Chateauroux the following day (fig.29) the Spanish air is evidently present above 800 mb, but here, on the right of the plume from Spain and where it ascends more slowly, the air is rather dry and there is no altocumulus. At Bordeaux (fig.30) the base of the Spanish air is at about 850 mb and it is moister, with a condensation level of about 700 mb, but the air a little higher is drier with a condensation level nearer 650 mb. On the isentropic chart for 37°C (fig.31) the air reaches this level over the English Channel, and consistently an altocumulus layer was reported over southwest England with bases at about 600 mb and cumuliiform tops reaching 450 mb in places. The θ_s in this layer was between 17 and 18°C.

c) The flow in the lower troposphere.

Over France on the 8th and 9th the screen-level potential temperatures reached values between 32 and 34°C (figs.26, 27). On the afternoon of the 8th high values of θ_w , reaching a maximum of 23°C, are found over western France in the characteristic disposition under the plume of air from Spain. The same pattern recurs on the

afternoon of the 9th. At Paris, somewhat east of the edge of the Spanish plume, the low-level air had arrived from the Mediterranean after passing over the Massif Central during the night hours, and was unusually dry (fig.32).

The isentropic chart for 31°C (fig.33) shows a marked distortion in the flow pattern north of Spain, which probably arises 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 moves from the southeast beneath the Spanish plume, and over the Bay of Biscay passes under its western boundary, marked approximately by the limiting streamline on the surface for $\theta = 37^{\circ}\text{C}$. Extensive castellanus formed over the central Bay of Biscay in a belt extending northwards towards the mouth of the English Channel, where adiabatic ascent of air with a θ_w exceeding about 18°C would provide buoyancy at all levels up to near the tropopause. Reconnaissance aircraft reported the castellanus bases between S.W. England and the 6th meridian over the Bay of Biscay to be between 8,000 and 10,000 ft, which with a θ of 31°C implies a θ_w of 17 to 19°C ; showers were observed as far west as 8°W in the mouth of the English Channel, well beyond the limiting streamline of the Spanish plume at about 5°W on the isentropic chart, and at about 4°W , across the Brittany peninsula, in its true geographical position. Sferics were located in the southern Bay of Biscay between 5 and 6°W on the evening of the 8th, and more extensively in a belt extending from 100-200 miles west of Bordeaux (3 to 4°W) to the Channel Islands (3°W), when sferic observations were resumed in the early morning of the 9th. A thunderstorm was reported at Brest ($4\frac{1}{2}^{\circ}\text{W}$) between 0700 and 0800, and at 0800 an intense cluster of sferics was observed just north of the Brittany peninsula. This evidently became the severe storm

which later reached Wokingham, for by 0900 the cluster had moved northeast into a position coincident with radar echoes which were subsequently followed continuously as they moved into and across S.E. England (Browning and Ludlam, 1962). This intensification into a severe storm occurred close to the position of the limiting streamline of the Spanish air, and just where the air at lower levels arrived from the region of the pronounced maximum in the screen-level θ_w on the previous afternoon (fig.26).

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 about 20°C (figs.34, 35). By mid-afternoon the main storm was decaying over the coast of East Anglia, but fresh intense storms developed over the main squall front successively farther south, first off the mouth of the Thames and later over Holland and Belgium (fig.36), where the adiabatic layer inland during the afternoon also acquired high values of θ_w (fig.27). These storms did not last long, however, dying out near the German border where even the screen-level θ_w fell below 20°C and where, well to the east of the frontal zone, the temperatures in the middle troposphere were higher (θ_s about 20°C). However, as is indicated as likely by fig.27, a further group of storms broke out in the evening near the coast north of Bordeaux and moved northeastward on a path similar but a little farther east than that followed by the previous storms, corresponding to a slight progression of the frontal zone and displacement of the Spanish plume. That the outbreak occurred in the evening rather than overnight or on the following morning, as in the previous case, can be related to the smaller distance which had to be traversed by the air modified over west France during the day, before it reached the western edge of the Spanish plume and the castellanous could develop.

d) Conclusions.

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

In both cases the storms developed in a frontal zone where there was the wind shear required to organise efficient cumulonimbus convection. The instability was associated with the production of an abnormally high Θ_w in the adiabatic layer over France, where in a rather narrow belt the convection was confined to a shallow layer beneath a plume of potentially warmer air from Spain. Showers occurred from castellanus clouds where the air from the adiabatic layer arrived at the western border of the Spanish plume and ascended, and from these a thunderstorm developed which became organised with a squall front. The thunderstorm travelled away from this region, in a direction somewhat to the right of the general winds, maintained by lifting over the squall front. It persisted for several hours until it had traversed the region occupied by the belt of high values of Θ_w . Since Θ_s in the middle and upper troposphere in the frontal zone near England is restricted to 17-19°C by the properties of the air in the flows which enter it from the south, the most significant and easily recognised feature of the situations is the production of a mean Θ_w exceeding about 19°C in the adiabatic layer over France, corresponding to a screen-level maximum Θ_w of 21 to 23°C.

3. The case of 29 August 1961.

The summer of 1961 was notable for the absence of thundery weather over England and France. Towards the end of the season a favourable large-scale situation arose, when a pronounced cold front approached the western seaboard, associated with a rather deep cyclone west of Ireland (central pressure about 975 mb on the morning of 29th August). During the 29th the cyclone moved northwest towards Iceland, while the front, and a deep associated trough on the conventional maps of the flow aloft, moved eastwards into the British Isles at a speed of 20 kts. At the time, on the morning of the 29th, thunderstorms seemed likely to occur in the vicinity of the front when it reached southeast England in the early evening, but in the event the front was accompanied by only a narrow band of altocumulus castellanus and some slight showers over the Welsh mountains, and in southeast England the sky remained cloudless. The situation was analysed to find an explanation for this behaviour, so different from that in the two previous case-studies.

a) The flow in the upper troposphere.

In the construction of the isentropic relative-flow charts the speed of the system was considered to diminish north of the 60th and south of the 35th parallel, becoming 10 kts at 65 and 25°N, and zero south of 20°N.

On the chart for the surface with $\theta = 43^{\circ}\text{C}$ (fig.37) prepared for mid-day on the 29th, the limiting streamline for the flow of the upper part of the Saharan adiabatic layer passes just west of the Canary Islands and Madeira northwards into the west coast of Ireland. On the sounding made in the Canary Islands (fig.38) the Saharan air has a base at about 850 mb and a θ_w of nearly 18°C; the top of the layer is between 500 and 450 mb and the lapse-rate there seems to have been modified by condensation. Very similar soundings were made at Kenitra and Gibraltar (fig.39),

and consistently small amounts of altocumulus, with some castellanus, were reported in the Canaries and thence northwards, and northeastwards into Portugal and southern Spain. On a slightly higher isentropic surface, more appropriate to the layer with clouds, the limiting trajectory leads back to south of the 20th parallel. In the west of the British Isles the Saharan air arrived at levels between about 550 and 400 mb (temperatures between about -15 and -25°C), accompanied by a rather narrow belt of high altocumulus, cirrocumulus and cirrus which increased to a complete cover north of about 55°N ; θ_s in the middle and upper troposphere in the frontal zone therefore had about the normal value of 18°C .

b) The flow in the lower troposphere.

Fig.40 shows that inland over Spain the screen-level maximum potential temperatures reached about 38°C on the afternoon of the 29th. Cumulus were reported over the higher ground, and the mid-day sounding at Zaragoza (fig.40) shows that their tops are unlikely to have reached above 600 mb and that the θ_w in the upper part of the convection layer attained its usual value of about 18°C . The isentropic chart for $\theta = 38^{\circ}\text{C}$ (fig.42) shows that this air arrived in the frontal zone over western England at about 600 mb (cf. the sounding at Camborne, fig.43) and beyond 50°N produced a layer of altocumulus in which θ_s was between 17 and 18°C . Small castellanus were reported in places, and a slight shower was observed by radar during the afternoon over the Welsh mountains, but no precipitation was observed by reporting stations over England and Wales.

Up to this point the structure of the air streams in the vicinity of the frontal zone does not seem significantly different from that in the two previous case-studies. However, when the air in the adiabatic layer over France is examined (fig.40) it is striking that in the west, beneath the plume of air from Spain,

the values of θ at screen-level are 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 have the remarkably low values of 4 to 6°C , corresponding to relative humidities at screen-level of less than 15%.

According to the discussion in Chapter II this anomaly indicates 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. This seems to be confirmed by fig.44, which shows that over the west and north of France the rainfall during the month was unusually scanty, being less than 50% of the normal over a large area and as little as 10% of normal (about 60 mm) in many places. The abnormally high potential temperature over Spain, and particularly over France, was achieved at the expense of the latent heat of evaporation. Moreover, in the west of France, and over Spain, 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 therefore seems reasonable that over large areas the soil had become too dry to provide considerable evaporation. In contrast, during the August before the Horsham storm (5 September 1958), the rainfall over practically the whole of France had been well above normal, by 50 to 100% over most of the west and north. 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 (fig.45) show the Spanish air present from 850 to 700 mb with a θ of between 36 and 37°C . Probably during the following afternoon the adiabatic layer over western France became hardly distinguishable from the Spanish air, the convection extending well above the 850 mb level where in the previous case-studies the plume from Spain provided a lid. On the morning sounding at Chateauroux (fig.46) the constancy of the mixing ratio up to 870 mb suggests that the convection had not yet extended above that level. At this time θ_w throughout the layer

below exceeded 20°C , but later the screen-level θ rose to 36°C and the convection in this region probably extended up through the Spanish plume to near the 700 mb level. As a result the mean mixing ratio in the bottom km decreased. Even at screen-level θ_w did not remain above 20°C , and the dryness of the air was such that θ_w higher within the adiabatic layer was generally only 18°C or less, and the condensation level was 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 (fig.47). Accordingly the only clouds observed during the passage of the frontal zone across S.E. England in the evening were the traces of altocumulus at 15,000 ft within the moister parts of the Spanish plume.

4. Severe local storms in the U.S.A. : the Geary storm.

a) General.

Severe thunderstorms, often accompanied by giant hail and tornadoes, have a greater frequency in the mid-western United States than anywhere else in the world, with a maximum in the spring and early summer in the states of Oklahoma, Kansas and Arkansas. At this season 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 in the lowest one or two km. Over the Gulf of Mexico in April the sea temperature is about 25°C , and θ_w is between 21 and 22°C in the adiabatic layer, which is usually rather less than 1 km deep. At the same time the mean values of θ_s in the middle troposphere over the states mentioned, at about 35°N , are as low as 16 to 18°C . The instability which leads to thunderstorm production is therefore not so much associated with recent sunshine overland, as in western Europe, but more with a persistent large-scale flow. Nevertheless here, as elsewhere, the worst storms break out in favourable synoptic situations, especially on the forward side of cold troughs approaching from the west. Moreover, 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 cold 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 at about the 800 mb level. Above the inversion θ is rather more than 40°C , and the lapse-rate very steep, but the inversion effectively prevents small-scale convection overland from distributing the moisture into a deeper layer. Tall

cumulus and cumulonimbus cannot develop until the inversion is removed, and a large part of the discussion in the literature is concerned with the possible mechanisms for its destruction. The lifting of deep columns through a pressure range of 100 to 200 mb is that most favoured, and it is suspected that this occurs predominantly (apart from that over the squall fronts of existing cumulonimbus) in the vicinity of the curious "dry-line". This is an almost north-south line across which during the day there is a very rapid transition at screen-level from the dew-points of 18°C or more which 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. This stream is considerably warmer, but there is practically no variation of virtual temperature across the dry-line. It is not regarded as a front ; usually as the whole 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 description of the processes at work there has not yet been given.

In order to examine the application of the isentropic relative-flow analysis to thunderstorm situations in this region, case studies were 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 Geary storm was one of a group of very severe thunderstorms which moved across Oklahoma during 4 May 1961. There were numerous tornadoes, falls of giant hail, and extremely heavy rainfalls. Radar echo tops were observed which imply that the cloud tops reached a maximum height of a little over 16 km (about the 100 mb level), some km above the tropopause. Browning and Donaldson (1963) have described the structure of the Geary storm in some detail, showing a close resemblance to the model of the Wokingham storm.

On the previous evening a squall-line had passed eastward across Oklahoma and Texas, and by the early morning of the 4th had moved into Missouri and Arkansas. The earliest showers associated with the severe storms over Oklahoma later in the day were reported over eastern Colorado a little before midnight on the 3rd, and in the following few hours spread across the border into Kansas and extreme northeastern New Mexico. By dawn thunderstorms were in progress just east of Amarillo, and soon afterwards these storms became severe along the border between Oklahoma and the Texas panhandle. The severe storms subsequently were confined to the southern fringe of a squall-line that extended from Nebraska to Texas ; although the squall-line advanced into neighbouring states no severe phenomena were reported after mid-evening. Until the 8th similar outbreaks recurred at approximately daily intervals.

Fig. 48 and 49 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 is not well marked, but the trough-ridge system appeared to be moving eastward at an average speed of 13 kts between the 3rd and the early hours of the 5th, although late on the 4th it slowed down considerably.

Similarly the dry-line shown over western Texas advanced eastward at about 13 kts before becoming stationary during the night of the 4th, and accordingly this has been taken as the speed of the system in the preparation of the isentropic charts.

c) The flow in the upper 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-scale convection. On the isentropic chart for 60°C (fig.50) the flow over the Great Plains appears to come westwards from the Pacific and there is no suggestion of any partition into distinct streams such as appears on lower isentropic surfaces. At most of the stations in the central and southern states the tropopause is found near the 250 mb level and with a θ close to 60°C (θ_s about 20°C). Accordingly during the intense cumulonimbus convection the anvil outflows can be expected to appear on or near this surface. Consistently, in the data of Browning and Donaldson (1963) the tops of the anvil echo more than 20 miles from the cumulonimbus dome of the Geary storm appear to be at about 200 mb, with the bottom of the anvil at 300 mb. It is interesting that on fig.50 there is extensive high cloud in a belt extending downwind from the storms over and near Oklahoma ; much of the other cirrus farther north and northwest may have its origin in smaller and isolated cumulonimbus which occurred over the Rocky Mountains.

d) The flow in the middle troposphere.

In the middle troposphere ahead of troughs over 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 equivalent regions are the arid lands of the southwestern states and of Mexico.

Most of Mexico is between one and three km above sea level. The mountains bordering the central plateau, particularly the Sierra Madre in the west, rise steeply from the narrow strip of coastal plain along the Pacific coast and from the rather broader Gulf coast plain in the east. In the spring the daily fluctuations of temperature and rainfall are small ; over the central area the afternoon screen-level potential temperature is mainly dependent upon height, and mostly lies between 44 and 48°C (fig.10). Over northern Mexico the monthly rainfall is invariably only a few mm from February until the monsoon is established in June, and, as in Spain and the Sahara, over such arid country the screen-level temperature gives a good indication of the potential temperature characteristic of a deep layer with an approximately dry-adiabatic lapse-rate. In the southwesterly winds prevailing ahead of troughs this layer is found aloft downstream with a potential temperature of about 46°C.

Fig. 51 shows a sounding made on the evening of 4 May over southern Texas on which such a layer is present ; its top is at 550 mb with a potential temperature of 45°C and a lifting condensation level of 410 mb, implying a θ_w of about 16½°C. Below the base of the layer at 750 mb is the cooler but moister and therefore potentially warmer layer ($\theta_w = 22^\circ\text{C}$ throughout) which has entered Texas as a SSE'ly flow from the Gulf of Mexico. This kind of sounding has long been recognised as a precursor of severe local storms in the American mid-west (Type 1 of Fawbush and Miller, 1954). Another notable feature of the sounding is the previously remarked tendency for a wind speed

maximum to occur in the upper part of the adiabatic layer aloft.

On the isentropic relative flow chart for $\theta = 46^{\circ}\text{C}$ (fig.52) the air stream which enters Mexico from the east is seen to turn northwards into a confluence ahead of the trough, and to have a limiting streamline which passes northeastward through a position very close to Oklahoma City. West of this streamline is air which has arrived from the west after flowing from the rear of the trough. The base of the Mexican air, with a potential temperature somewhat below 46°C , is very sharply defined over southern Texas at levels between 800 and 700 mb, and the air can be recognised by a very steep lapse-rate in a layer about 200 mb deep. 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. Over and downwind from Oklahoma the flow of Mexican air is difficult to identify, probably because of modification in the thunderstorm region. Generally the Mexican air was too dry to produce cloud during its flow over the southern United States, other than small amounts of alto-cumulus in a rather narrow belt originating over southern Texas (fig.52). According to aircraft reports the height of this cloud was between about 550 and 420 mb, implying the rather high θ_w of about 18°C likely to be found only downwind of intermediate scale circulations associated with the mountains rising above the general Mexican plateau level.

The relative flow chart for the surface $\theta = 37^{\circ}\text{C}$ (fig.53) shows the flow into the confluence zone of the 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 to the flow from the Caribbean appears to pass just west of Oklahoma City and constitutes the dry-line at about the 700 mb level. In this flow the air is rather moist, with a lifting condensation level of about 600 mb, but west of the dry line is encountered the

opposing flow, which after subsiding in the rear of the trough passed across the desert lands of the southwestern United States and northern Mexico. There the screen-level potential temperatures just reached 40°C and a deep dry-adiabatic layer with a potential temperature of about 37°C was produced. Since the flow within a deep layer to 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 very dry, with a mean θ_w of only about $14\frac{1}{2}^{\circ}\text{C}$. The modified air is generally described as "continental tropical", but it reaches the confluence zone from middle latitudes. Its properties on this occasion are well illustrated by the sounding from Amarillo (fig.54), about 150 km west of the dry-line. The air was too dry for any cumulus to form in the convection layer, except perhaps locally over mountain peaks.

There is a general similarity between the flow pattern in and around the confluence zone and those previously described in the European case-studies. A striking difference, however, is that in the latter the flow which enters the trough from the west becomes subject to small-scale convection mainly over an ocean whose surface temperature is only between about 15 and 20°C . Consequently this flow remains cold, in contrast to the flow east of the confluence line, which passes over ground where surface temperatures in the range 30 to 45°C are attained. Consequently over western Europe the confluence line is recognised as lying in a rather intense cold front zone, and the flow on its eastern side is even at low levels subject to a large-scale baroclinic ascent which has been shown to play an important part in releasing the instability. Over the southwestern 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 preferred position is probably strongly related to the topography of northern Mexico and the southern Rockies. The gap between

these high lands, in the vicinity of the northern border of Mexico, 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 the Chihuahua region of northern Mexico.

e) The flow in the low-troposphere.

The flow across the coast of the Gulf of Mexico into the southern United States which is indicated in figs. 48 (for the 1000 mb level) and 53 (for about the 700 mb level) consists of air which has been modified by small-scale convection over the ocean in the Atlantic trade winds. On arrival in the Caribbean the potential temperature in a rather shallow adiabatic layer has become very nearly the sea surface temperature, about 26°C in May. At deck-level θ_w is about 23°C ; slightly lower values are found in the adiabatic layer, and values of 19 to 20°C in the layer occupied by the trade wind cumulus which extend to about the 800 mb level (see, for example, the sounding from Swan Island in fig. 55. The entire 200 mb layer below 800 mb is known as the trade wind moist layer; the air above, up to the top of the easterlies at middle levels is considerably drier.) Consequently the mean value of θ_w in the lowest km or two in the flow which enters the southern states is in early May between 21 and 22°C . Over the Caribbean generally θ_s in the middle troposphere is about the same (fig. 49), but its value decreases northwards and on this occasion was only about 18°C or less inland. Ordinarily it could be anticipated that as the trade wind air moved inland and became subject to further heating while the mid-tropospheric temperatures decreased, the small-scale convection would intensify and extend upwars with the widespread development of

showers and thunderstorms. The potential instability would thus be dissipated over a large area, probably without any violent effects.

However, inland over Texas the air lay below a lid formed by the plume of potentially warm air downwind of Mexico. The base of this plume, whose inclination is indicated on fig.52, limited the small-scale convection to the lowest 200 or 300 mb. Accordingly, as the air flowed inland the depth of the moist layer 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 (fig.56) to 22°C where the air reached northern Texas after a period of a day or more over land (fig.57). On the chart for $\theta = 32^{\circ}\text{C}$ (fig.58), which over western Texas and New Mexico is blank in the presence of a higher θ in the adiabatic layer, the flow can be followed approximately as far as latitude 35°N and then disappears from the isentropic surface, the air evidently being transported to the high troposphere in cumulonimbus ; beyond Oklahoma City the values of θ_w everywhere in the troposphere are mainly less than 18°C , and only at the tropopause was θ_w as high as 22°C . Nearly all the severe storms appear to have originated near the dry-line, and almost all the reported severe incidents, when adjusted to their correct relative position, are at distances of between 30 and 120 mi of it, and within a region of destabilization shown on fig.16 where the moist flow, continuing northward along a general path east of the Sierra Madre mountains and parallel to the mountain chain of North America, reached the edge of the Mexican plume near the dry-line ; beyond the Mexican plume there was no longer any restraint to deep convection. The general confluence of the streamlines approaching Oklahoma is augmented somewhat by a slight component of flow toward the dry-line, probably associated with the cumulus convection all along it, which imposes a net convergence in the lowest few hundred metres. The progressive

development of instability is illustrated by the soundings from near the coast (fig.56) where the convection is confined below 820 mb ; from central Texas (fig.57), where the convection reaches 700 mb (and the Mexican lid is somewhat less restrictive to the convection) ; and finally from the neighbourhood of the dry-line beyond the Mexican plume (fig.59). Here the Mexican air is not present aloft and the moist air is evidently intruding below the potentially warmer (in respect of unsaturated processes) air of the deep dry-adiabatic layer of continental tropical air, produced over the desert farther west. In this vicinity cumulus congestus formed from the moist layer, with a base probably between 1500 and 2000 metres above the ground, are readily able to grow through the adiabatic layer and for some distance above its top at about 550 mb, probably becoming large enough to produce showers and thus to initiate cumulonimbus such as were reported as within sight of Altus at 1730 C.S.T.

However, the severe storms were not dependent upon initiation near the dry-line this afternoon, for cumulonimbus had been developing into an extensive squall-line since the early hours of the morning. Late on the previous evening there were showers in S.E. Colorado which had probably formed over the mountains in the afternoon, and these groups moved east with fresh development on their southern flank to produce a squall-line which by 0400 was more than 200 miles long and extending into northern Texas near Amarillo with radar echo tops to 44 to 46,000 ft.

It is interesting that on the afternoon and evening of the 3rd isolated cumulonimbus were observed by radar over the eastern slopes of the Sierra Madre mountains west of Brownsville and thence northwards in the vicinity of the limiting stream-line of the moist low-level flow. These cumulonimbus were evidently unable to propagate away from the slopes where they originated, probably because the energy of the downdraughts was

insufficient to raise the moist air through the lid formed by the Mexican plume, which over western Texas away from the dry-line was a strong obstacle. Similar behaviour is discussed in the following case-study.

At Oklahoma City the sounding (fig.60) made just east of the Geary storm shows a stratification which, while not able to produce large cumulus, was by virtue of the steep lapse and low θ_s in middle troposphere, very favourable for the propagation of the approaching squall-line. At the time of the sounding there was an overcast of high cloud, probably the anvil cloud of the severe storms, whose base appears to be marked on the sounding at about 300 mb, where there is a distinct rise in the humidity and in θ_s , the latter attaining the value of 21 to 22°C consistent with the values of θ_w in the lowest km. The adiabatic ascent of this air implies temperature excesses of as much as 7°C throughout the middle troposphere and accordingly a greater buoyancy than is normally attained in European storms, but about the same as that in the exceptionally severe Horsham storm discussed earlier. Values of buoyancy were even greater in the region of destabilization at and beyond the edge of the Mexican plume (see fig.59). The intrusion of the moist layer beneath the continental tropical air near the dry-line and away from the region with a strong lid is indicated on the composite chart (fig.61). It also shows the formation of the dry-line by the confluence of air streams from the trade wind origin and from the rear of the trough.

f) Hydrological aspects.

The continuous intrusion of trade wind air into middle latitudes, which is brought about by the interruption of the sub-tropical easterlies by a high land mass, results in a substantial flux of water vapour in a stream of 200 mb depth

and 500 miles width along the south coast of the United States. On the basis of fig.51 the flux of water vapour in this trade wind moist layer is 11×10^{14} g/hr (henceforth referred to as 11 units). The average rate of rainfall during the 4th was found by integrating the falls over the entire area of the states south of latitude 39° , using data from a dense network of hourly reporting stations, and was 7 units. Since the thunderstorms were undoubtedly organized, the ascent of all the trade wind moist layer to the vicinity of the tropopause must be anticipated to have been accompanied by downdraughts which replace the original air flow. In the downdraughts a considerable fraction of the condensed water is evaporated in and below the cumulonimbus in air originating in middle levels, and the appearance of this rain-cooled air at the ground is often manifested at the height of the activity by an extensive high-pressure area several hundred miles in diameter behind the squall-line (Fujita, 1959). Regardless of the degree of saturation in the downdraught, the θ_w within it will be that in the middle troposphere in the vicinity of the thunderstorms, though its potential temperature will be somewhat higher than in the cold air behind the warm front. The Topeka sounding (fig.62) shows this outflow air as a nearly saturated layer between 830 and 630 mb, with a θ_w of 16°C throughout. Behind the squall-line and generally above the cold air which had a θ_w of only 10°C below 830 mb (fig.62), the isentropic chart for 32°C (fig.58) describes the flow of air which replaces the trade wind moist layer. On this chart is indicated an extensive belt of altocumulus stretching from Oklahoma to the Canadian border, parallel to the streamlines which lead away from the region of thunderstorms. Aircraft reconnaissance observations confirm that the 200 mb deep layer below 630 mb at Topeka does contain a dense cover of cloud.

The width of the altocumulus belt which can definitely be ascribed to upslope motion in the outflow air is 350 miles at

39°N, the latitude of Topeka. According to calculations based upon the Topeka sounding (fig.62) the flux of water vapour across the 39th parallel in the altocumulus layer is 4 units. The difference of 7 units between this and the flux in the moist layer entering the Continent agrees with the measured precipitation south of Topeka, some of which undoubtedly came from the altocumulus and not from the thunderstorms. Assuming that the air between 830 and 630 mb at Topeka had ascended from a position at the ground south of the warm front and was initially saturated, 3 units might have fallen from the altocumulus. This implies that as much as 7 units in all are carried away from the thunderstorms as re-evaporated moisture, most of which would fall eventually as showers farther upstream. A negligible amount of precipitation is re-evaporated into the initially dry easterly current flowing under the altocumulus, but resulted in the formation of a layer of stratiform cloud beneath the warm front inversion (cf. the saturated air from 830 mb to the ground in fig.62). It has also been assumed that the amount of moisture flux in the cumulonimbus anvils in the high troposphere is negligible.

Browning and Donaldson (1963) have shown that a depth of about 120 mb of potentially warm air entered the Geary cumulonimbus on a flank about 20 miles across, implying a water vapour flux of one-half unit into the base of the cloud. Their calculation, therefore, suggests that twenty cumulonimbus of the size of the Geary storm would be required to remove all the water vapour from the whole trade wind stream, 500 miles across. Since there were never more than about 6 such large cumulonimbus present simultaneously, even at the height of activity, it is likely, in view of the agreement between measured and calculated rainfall rates, that only a small fraction of the total precipitation fell from the storms of greatest severity.

g) Seasonal values of Θ and Θ_w

The northward deflection of the trades in the spring and the tendency for troughs to become stationary west of the continental divide are geographical effects. The establishment of a characteristic large-scale trough-ridge pattern strongly related to topography and to fixed heat sources in low latitudes suggests that both Θ and Θ_w in the characteristic air streams are unlikely to vary appreciably from one situation to another. Table 7 contains an estimate of the mean Θ_w in the bottom 100 mb of the trade wind air, the mean Θ and Θ_w in the Mexican air, and the mean Θ and Θ_w in the continental tropical air, based on data from 1961. The observations used were arbitrarily chosen so that certain obviously inappropriate soundings were excluded.

TABLE 7

	early <u>April</u>	late <u>April</u>	early <u>May</u>	late <u>May</u>	early <u>June</u>
<u>Trade wind</u> <u>air</u>					
Caribbean	22.2	21.7	21.9	22.5	22.4
Texas (under lid)	18.2	21.6	22.0	23.1	22.0
<u>Mexican air</u>	44.5 16.4	45.5 17.5	45.5 17.6	46.5 17.2	45.5 16.8
<u>Continental-</u> <u>tropical air</u>	33.0 13.4	37.5 14.6	37.5 14.8	39.0 17.2	42.0 17.6

The Table shows that the Θ_w in the trade wind air over Texas remains practically constant after early April and is the same as the Θ_w computed at a station in the Caribbean. There is a slight progressive warming apparent in the Mexican air, and a rather rapid

one in the continental tropical air, so that the two streams do not differ greatly with respect to potential temperature by June, at the time of the Mexican monsoon.

5. The case of 26 May 1962.

On 26 May, 1962, numerous intense thunderstorms occurred over Texas, Oklahoma, Kansas, and western Missouri, and there were many reports of tornadoes, extremely heavy rainfall, and giant hail. In one location hail covered the ground to a depth of six inches, and in another the maximum stone diameter was about four inches. Particularly in the evening, when the squall line was fully developed, the non-tornadic wind gusts became destructive near its southern edge. Severe storms had occurred also throughout the area on 25 May, and continued to recur for several days after that.

In figs. 63 and 64 the surface and upper air patterns on 26 May appear similar to those on the day of the Geary storm except in that the warm front was situated farther north. The dry-line extended from Kansas across Texas, and does not appear to have been moving much in any direction. The eastward movement of the large-scale trough-ridge system averaged over 48 hours was about 5 knots, and this speed has been chosen to obtain the relative winds on the isentropic charts.

a) The flow in the middle troposphere.

Table 7 suggests that mean values of θ and θ_w in the deep convective layer over the high Mexican plateau are about 47° and 17°C during late May. The Del Rio sounding (fig.65) shows this characteristic adiabatic layer between 510 and about 790 mb with values of θ and θ_w indeed very close to those listed in the table. One peculiarity on the sounding is a layer of moist air above 510 mb, probably produced by condensation over Mexican hills, and consistent with the presence of extensive altocumulus and low

cirrus over southern Texas. There was a pronounced increase in wind speed with height in the Mexican adiabatic layer and a wind speed maximum at the top of it.

The diagram constructed for the 47°C isentropic surface (fig.66) demonstrates the similarity with the previous case (see fig.52). East of the limiting streamline, the top of the Mexican plume ascended from an estimated 550 mb over the plateau to 450 mb in the region of thunderstorms. The greatest total ascent and the strongest vertical velocities were, as to be expected, nearest the front. Condensation levels in the Mexican air varied between 390 and 450 mb and so large-scale condensation at the top of the layer did not occur except in the moist plume which appears in both cases as a notable band of low cirrus and altocumulus emanating along the streamlines from the northeastern part of Mexico.

In the absence of interference by thunderstorms, the characteristics of Mexican air persist recognizably for a long time. Fig.67 shows that at Charleston, about 1000 miles downstream from Mexico, the Mexican plume is present between 680 and 520 mb, with a wind speed maximum near the top. Since the top of the layer has a potential temperature of 44° (which is 40 mb below the 47°C surface) and the mixing ratio is practically identical to that within the Mexican plume at Del Rio, a cooling of 3°C , presumably by radiation, must have transpired in the two days since the air left the convective layer over Mexico.

As shown in the previous chapter, the boundary between the Mexican air and the flow rounding the trough forms the middle-level extension of the dry-line. On 26 May the layer of small-scale convection west of the dry-line was especially warm and deep, and at Amarillo (fig.68) the adiabatic layer extended up to 520 mb, below which the mean values of θ and θ_w were $42\frac{1}{2}^{\circ}\text{C}$ and 16°C respectively. The condensation level at the top of the layer was about 425 mb.

The isentropic chart showing the relative flow on the $42\frac{1}{2}^{\circ}\text{C}$ surface is presented in fig.69. The chart illustrates that the adiabatic layer west of the dry-line was formed in a current of air which had previously descended behind the trough before flowing for one or two days over the moderately high and arid terrain of northwestern Mexico. Air in streamlines passing west of Albuquerque had spent less time over arid ground and the air in the adiabatic layer over Arizona was accordingly less warm. Condensation did not occur in the continental tropical air because it never ascended very far above 520 mb. As in the Geary case, the sounding at Amarillo is typical of a region about 400 miles wide west of the dry-line.

Continental tropical air can be recognized on the Columbia sounding (fig.70), made hundreds of miles downstream from Amarillo. The characteristic very dry quasi-adiabatic layer was there located between 520 and about 740 mb and had a θ_w of 15°C throughout. The continental tropical air leaves the ground at the surface position of the warm front ; the bottom of the adiabatic layer moves upwards along a slanting isentropic surface and arrives downstream above the ground at a height equal to the depth of the cooler air below. Where the flow was over the trade wind air (present below 770 mb on the Columbia sounding) it formed a barrier to the convection which was far less restrictive to convection than that produced by Mexican air but sufficient to ensure that in the vicinity of Columbia the cumulus were small and scattered.

b) The flow in the lower troposphere.

Screen-level potential temperatures on 26 May were mainly between 28° and 35°C south of the warm front, rising to about 36°C near the dry-line. Fig.71 shows the relative flow on the isentropic surface for $\theta = 35^{\circ}\text{C}$. On most soundings in the trade air, such as that from Del Rio (fig.63), this surface is

slightly above the top of the adiabatic layer. Included in fig.69 are isopleths of the mean θ_w in the lowest 100 mb of the trade wind air. The diagram shows that the moist trade wind stream turned northward ahead of the Mexican plateau, so that the dry-line defining its western boundary became the northward projection of the edge of the plateau. The mean θ_w in the lowest 100 mb between the Caribbean and Texas was 22° to 23°C , and the depth of the moist layer in the vicinity of the thunderstorms was about 200 mb, just as in the trades. As the trade air over the Gulf accelerates toward the north it appears to diminish in thickness, so that the depth of the moist layer along the northern part of the coastal plain was comparatively small, while beyond this region the isentropic surfaces slope upward toward the north.

Fig.71 indicates that the 35°C isentropic surface slopes upward through the region of thunderstorms. In the previous chapter it was concluded that the cool and moist thunderstorm downdraught moved northward, replacing the trade wind air. The area of middle-level cloud north of the warm front lies roughly within the streamlines that form the continuation of the trade wind flow. Analogous to the Topeka sounding (fig.62) in the previous chapter is the one made at St.Cloud on 26 May (fig.72). The altocumulus there corresponds to the practically saturated layer which can be seen on the diagram to coincide between 700 and 560 mb with the saturated adiabatic for $\theta_w = 16^{\circ}\text{C}$. The particular wet-bulb potential temperature of 16°C in the altocumulus layer is consistent with that to be expected in the air descending from middle-levels in cumulonimbus downdraughts: θ_w there in the vicinity of the thunderstorms was between 15° and 17°C . Soundings similar to that at St.Cloud were made within the cloud system at Green Bay and Flint. Substantial large-scale downward motion near the east coast resulted in the evaporation of the cloud there.

c) The composite flow pattern.

The composite flow chart (fig.73) for 26 May is similar to that for 4 May (fig.61). On each occasion in a region where the moist layer passed beneath the interface between the Mexican and continental tropical flow near the dry-line, the atmosphere was destabilized (with respect to cumulus convection) by a kind of differential advection. Most of the severe storm activity on 26 May was associated with cumulonimbus which originated within this region and produced severe incidents in a zone between 30 and 120 miles from the dry-line and within 30 miles of the interface between the Mexican and continental tropical air. As in the previous case little rain fell south of latitude 33°N , where there was a strong lid.

Analogous to the 4 May Altus sounding (fig.59) is the one made at Topeka on 26 May (fig.74). This station was located in the destabilized region and very close to severe thunderstorms. The depth of the moist layer was only 200 mb, not much greater than it was upstream in central Texas, indicating only modest large-scale ascent in the interval. No pronounced inversion existed at the top of the moist layer there and the continental tropical air present above 760 mb was decidedly cooler than the Mexican plume located a short distance farther south. There were scattered cumulus near Topeka, but some distant cumulonimbus were in sight just as at Altus on 4 May.

Generally the mean θ_w in the bottom 100 mb of the moist layer was about $22^{\circ}\text{--}23^{\circ}\text{C}$ and aloft in the Mexican air it was 17°C . Although in middle-levels the extension of the dry-line represented a kind of front, it was not there a region of significant temperature discontinuity. In the top of the Mexican air the relative humidity was rather high and, accordingly, a short distance away in the destabilized region θ_s could not much exceed 17°C . (See the schematic cross-section in fig.75.) The combination of this value of θ_s and the θ_w below produces a Showalter index of -7 to -9.

Although the horizontal temperature change across the dry-line is small below the level of the lid near 700 mb, and above 520 mb, there is considerable difference in temperature between the Mexican and continental tropical air between these levels. According to Table 7, in May the difference may be 8°C , corresponding to a difference in θ_s of about 5°C . The total positive area on a thermodynamic diagram is therefore enhanced considerably when continental tropical air replaces Mexican air above the moist layer, although a conventional stability index (using data from only the 500 and 850 mb levels) may not reveal this. The suggested structure is consistent with previous inferences that the dry front slopes upward to the east at low levels but bends backward somewhat towards the west above about 600 mb (see Fujita, 1958, p.582 ; McGuire, 1962, p.5). The zone of greatest instability (in respect of cumulus convection) is a little to the east of the surface position of the dry-line and very near or a little to the west of the limiting streamline in the Mexican air. These features are illustrated in the schematic cross-section of the dry-line drawn in fig.75.

d) The influence of insolation east of the dry-line.

It has been shown that if the introduction of the energy corresponding to the net radiation is confined to the bottom 150 mb, the increase in θ_w in the layer during a clear summer day is about $2\frac{1}{2}^{\circ}\text{C}$. In both the cases studied the isentropic surfaces east of the dry-line slope upward toward the north, so that the lid appears to have confined the small-scale convection to the bottom 200 or 250 mb above the ground over Oklahoma. The increase of the mean θ_w along streamlines in the moist layer arriving over Oklahoma, assuming clear skies and a net long wave loss from the layer equivalent to a cooling of $\frac{1}{2}^{\circ}\text{C}/\text{day}$, can therefore be anticipated to be about $1\frac{1}{2}^{\circ}\text{C}/\text{day}$.

The rather elevated base of the lid and the high humidity in the trade wind air result in a cloud base which is below the top of the moist layer. Numerous small cumulus appear just beneath the lid over land and the convection involves a mixing which produces a nearly saturated condition at the top of the adiabatic layer. Not only is incoming radiation reduced by the cloudiness, perhaps by as much as 25%, but the net outgoing radiation is enhanced by the high humidity. Often, as on 4 May, 1961, the amount of heat loss after sunset is sufficient to produce a stratus layer. This further enhances the long wave loss and encourages the thickening of the low cloud. Further ascent along isentropic surfaces during the night may produce very light precipitation and the stratus layer may continue to persist after sunrise. Computations with an Elsasser radiation diagram show that the long wave radiation loss at night may be an additional 1°C (corresponding to $\frac{1}{2}^{\circ}\text{C}$ in θ_w) when there is stratus formation. The final estimate for the increase in θ_w in the moist layer east of the dry-line is $\frac{1}{2}$ to $1^{\circ}\text{C}/\text{day}$ under normal cloud conditions, and about $1\frac{1}{2}^{\circ}\text{C}/\text{day}$ as a limit under totally clear skies. The increase in θ will be a fraction of a degree per day when the ground is moist but approach $3^{\circ}\text{C}/\text{day}$ over arid terrain.

The isentropic charts lead one to envisage that the easterly flow of the moist trade wind layer is turned into a southerly current upon confrontation with the high barrier of Mexico. The northernmost streamline in the trades becomes the eastermost streamline over the United States and the streamline lying along both the Sierra Madre mountains and the dry-line originates far south in the trades (at about latitude 15°N). Cumulus convection along the mountain slopes and the dry-line introduces some flux toward the west at very low levels. This is especially noticeable at the surface, where the winds often flow strongly toward the dry-line. The latitudinal increase of about 1°C in θ_w towards the south in the trades

imposes a similar increase from east to west where the flow enters the mainland. In the trades in May the depth of the layer of small-scale convection is normally about 200 mb ; the mean Θ_w in the bottom 100 mb is 1-2°C higher than in the rest of the moist layer, reflecting the fact that the cloud layer, which constitutes the upper two-thirds of the moist layer, is associated with an appreciable decrease with height in Θ_w . As soon as the heating is introduced during the flow over the North American mainland, the vigorous convection occupies nearly the whole depth of the moist layer beneath the lid, and the original properties of the trade wind air tend to become uniform throughout the new adiabatic layer. The convection may increase Θ_w by between ½°C and 1½°C/day averaged over the whole of the adiabatic layer, but it will not alter Θ_w in the bottom 100 mb by more than ½°C, and for all practical purposes Θ_w there is conserved along the flow.

Parcels following successive trajectories passing east of Brownsville in the moist layer remain in convective contact with the ground for a duration of about one day before arriving in the Oklahoma region. It is thought that the effects of nocturnal cooling and the distribution of cloudiness may produce a diurnal variation in Θ_w of about ¼°C in the flow arriving in the region of thunderstorms. Obviously the heating is of secondary importance in the production of the instability, although it will be shown presently that it can be responsible for the initiation of cumulonimbus. Whereas the lid at the base of the Spanish air over France produces instability by raising Θ_w below, the lid at the base of the Mexican air prevents a reduction of Θ_w in the lowest layer by the spread of the small-scale convection upwards, until the flow has progressed to a position where Θ_s aloft is several degrees lower.

In the spring, monthly precipitation decreases westward

across the central plains and amounts are extremely small over western Texas and the broad Mexican coastal plain. Air in the moist layer over eastern Texas receives a rather small fraction of the net radiation as sensible heat, and accordingly the mean θ over the bottom 200 mb increases very little. On 26 May, the screen-level θ and θ_w a great distance from the dry-line were 34° and $24\frac{1}{2}^\circ\text{C}$ respectively, and the cloud base, as defined by these quantities, was about 1700 metres (above M.S.L.), about the 1400 m estimated by photogrammetry (fig.76). Since the base of the lid was about 2500 m, vertical development of the cumulus at this distance from the dry-line was not appreciable, as can be seen in the photograph taken at 14,000 ft near Oklahoma City at 1345 C.S.T. (fig.77 ; see also fig.78).

Along trajectories passing near and west of Brownesville the duration of travel over land prior to arrival in the region of thunderstorms becomes appreciably greater than one day. Because sensible heating predominates along this part of the flow, there is a marked increase in potential temperature westwards within about a hundred miles of the dry-line. Conversely, fig.75 shows that the mixing ratio decreases westward in accord with the diminishing importance of energy introduced as latent heat. It is evident that the cloud base rises toward the west, since x decreases and θ increases (see fig.76), and low cloud disappears about where the lifting condensation level lies above the lid base. Fig.77 shows that on 26 May the system of small cumulus terminated near Oklahoma City ; farther west the only low cloud was close to the dry-line.

Immediately east of the dry-line the screen level θ of 39 to 40°C hardly differed from that in the dry air on its other side. It can be expected that over hills and other favourable ground features differential heating raised θ by a degree or two and gave rise to intermediate scale cir-

culations in which the cumulus formed and then readily grew to considerable height in the absence of the lid. At the dry-line and in a zone on the moist side widening along the direction of flow either the lid at the base of the Mexican air is very high and therefore weak, or the Mexican air has been replaced by the continental tropical air. This zone is highly unstable for cumulus convection, since there is an extremely steep lapse-rate over a deep layer above the ground and little to restrict their growth up to a height of at least 5 km near the top of the layer of continental tropical air.

The small-scale convection near the dry-line does lead to mixing between the trade wind and the continental tropical air, and accordingly in fig.76 there is distinct decrease of θ_w westwards from a position about 75 miles east of the surface position of the dry-line. In the intermediate zone the adiabatic lifting condensation level of the screen-level air rises steadily towards the west, and may generally be rather too high to support the formation and growth of vigorous cumulus. Ordinarily the observed base of cumulus is appreciably above the condensation level corresponding to screen temperatures (p.12), but remarkably estimates made from aerial photographs of the cumulus clusters near the dry-line and included in fig.76 indicate a base-level lower by about a km. Although these estimates are not very accurate they can be regarded as supporting the view that the clusters appear where favourable ground features induce intermediate-scale circulations in which the adiabatic layer is locally moister. These circulations may be strengthened by the release of latent heat when some of the cumulus produced develop towards the congestus stage, so that after a rather long period (several hours following sunrise) in which none or only small clouds are present, the transition to towering cumulus and shower formation may be comparatively rapid (within an hour : see figs. 83-85).

More pronounced intermediate-scale circulations occur over the foothills and peaks of the mountains, such as the Sierra del Burro near Del Rio and the Sierra Madre, along the dry-line farther south. On both 4 May 1961 and 26 May 1962 isolated and practically stationary radar echoes were detected in these regions during the afternoon, indicating the local development of tall cumulonimbus. Strikingly, these echoes disappear in the evening ; the cumulonimbus are evidently unable to move away from the mountains as organised systems, probably more because of the presence of the restrictive lid over the moist air than because of the absence of a favourable wind shear.

e) Observations of cumulonimbus formation near the dry-line.

The photographs used to estimate the cumulus base-level near the dry-line were from an extensive series made during the afternoon of 26 May 1962 on a research reconnaissance flight. Fig.78 shows the positions from which the photographs reproduced in figs. 77 and 79-86 were taken, and also the positions of the dry-line and severe storm events.

The first swelling cumulus near the dry-line were reported in a position 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., and are shown in full maturity in fig.79 at 1500 C.S.T. At this time its dome appears to be well above 40,000 ft, the height of the tropopause. A second cumulonimbus group (B ; fig.80) appears to have developed about an hour later from the southern part of the same source region. The subsequent pictures illustrate the rapid development of cumulonimbus (groups C, D, E) from widely separated cumulus clusters close to the dry-line. Although the spaces between these clusters are almost completely clear of cloud, in places (as between the cumulonimbus groups C and D in

fig.86) there are sufficient small clouds to encourage the view that the disposition of the dry-line can be seen visually.

An advanced state of glaciation was observed in congestus clouds which reached the altitude of the aircraft (400 mb), and anvils appeared when their tops began to exceed 35,000 ft (250 mb). Between these two stages, the precipitation presumably initiated downdraughts and minor squall fronts. The composite diagrams (figs. 61, 73) show that the wind veers strongly with height above the moist layer ; partly as a consequence of the flow at lower levels being highly influenced by the configuration of the terrain. Considering also the tendency for propagation over the squall fronts to occur on the right flanks of the cumulonimbus, it is clear that the individual clouds will move away from the dry-line and towards the region of maximum Θ_w at the surface. The marked wind shear is illustrated in fig.81 by the lean of the congestus away from the dry-line (whose orientation is shown by the line of small clouds about 10 miles from the congestus). The cumulonimbus group A in the background of this picture has already moved 25 miles away from the dry-line.

Congestus growth in the vicinity of the thunderstorms may be predominantly over the squall fronts, as in figs. 81 and 79, and not closely related to the ground heating. The cumulonimbus propagation will be encouraged everywhere within the region of destabilization, but strongly impeded by the lid where the storms reach the Mexican air, even though the wind shear is favourable and a large buoyancy can be attained following sufficient lifting of the moist air. Fig. 78 shows that 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. The propagating squall-line tends to seal off the flow of moist air from the south ; as it moves eastward into the Mexican air, away from the highest values of Θ_w near the ground and toward higher

values of θ_s aloft, it diminishes in intensity (although the rain continues to be quite heavy until its final dissolution). However, the persistence of the large-scale pattern enables the original situation to be re-established within a day, so that the same sequence of events may recur.

Conclusions.

The production of severe local storms requires the development of great buoyancy in the presence of wind shear ; the former can be expressed by the difference between θ_w near the ground (in the layer of small-scale convection) and θ_s in the middle and upper troposphere. It appears that the low-level θ_w is far more variable than θ_s aloft in a given part of the flow pattern and time of year, because the former is more immediately influenced by the terrain, whereas the latter is controlled mainly by the large-scale circulation. The rise of θ_w near the ground will lead to appreciably greater values in the bottom kilometre than those found at middle levels, if the convection due to sunshine absorption at the ground is confined to the bottom 1 or 2 km for a period of a day or two. The production of high buoyancy is favoured in the southerly flow ahead of a major trough, where the air-streams aloft cool by large-scale ascent while the small-scale convection may be confined beneath a thermal barrier (or 'lid') formed by the advection of a very warm adiabatic layer aloft from a hot and arid plateau.

In the case of thunderstorms over western Europe and the British Isles, severe outbreaks tend to occur in the vicinity of slowly-moving cold fronts and ahead of major troughs approaching from the Atlantic. When the heating proceeds for a day or two over the comparatively moist French plain and is confined to the bottom 1 or 2 km beneath a lid produced by the advection aloft of the very warm adiabatic layer from the Spanish table land, the rise in θ_w is significant. The warmth of the Spanish adiabatic layer is attributed to the aridity and high elevation of the Iberian peninsula. A pocket of high θ_w is therefore first observed in the lee of the Pyrenees and then 24 hours later over northern France. Thunderstorms break out north of the Channel, where the Spanish

air has ascended and cooled a sufficient amount to remove the restraint to the growth of cumulus. The first showers appear to begin at the edge of the Spanish plume and near the upwind edge of the altocumulus castellanus system formed by condensation in the greater body of the French adiabatic layer.

At middle and upper levels the flow ahead of the trough is derived from low latitudes, and important airstreams originate in the layers of small-scale convection over Spain and the Sahara. Aloft over western Europe these approach saturation and then θ_s is practically identical to the slowly-changing values of θ_w in these regions. Isentropic streamline charts illustrate a model of the flow pattern in which the airstreams travel from the region of small-scale convection in the south up surfaces of constant potential temperature inclined northward. The major cold front present is considered to be the air mass boundary and a surface through limiting streamlines at the edge of this ascending flow. The confluent frontal zone is a region where the necessary shear for severe cumulonimbus is present.

The rise in θ_w over France may not be sufficient to produce thunderstorms if through drought the soil lacks sufficient moisture. In that case θ may rise abnormally high and by approximating to that in the Spanish air remove the restraint on the convection.

Over the Great Plains of the United States the high θ_w near the ground, which is associated with exceptionally frequent and severe thunderstorms during April and May, is provided by the large-scale intrusion of the trade wind moist layer. While the moist layer flows northward parallel to the edge of the Mexican plateau prior to entering the United States, the easterly flow at middle-levels enters the high plateau, where intense heating in the virtual absence of evaporation produces a deep adiabatic layer. The plume of Mexican air is advected northward and it forms a lid aloft over the moist layer ; as

in the case of Spanish air over France. By restraining the convection to the bottom 2 or 3 km the lid serves to conserve the high θ_w in the trade wind moist layer (or permit it to increase slightly) while the Mexican air ascends to reach a state of near saturation at the top ; thus θ_s attained at middle-levels downstream from Mexico is practically equivalent to θ_w in the Mexican adiabatic layer and considerably less than θ_w in the moist layer.

The edge of the Mexican and trade wind flow forms a front, known as the dry-line and discernible below 400 mb, which separates the flow of this air from a very warm and dry current which has been heated mainly over desert regions following its descent during flow from the rear of a major trough. Near the dry-line the relatively cool trade wind air flows out from under the Mexican plume and increasingly undercuts the dry air to the west, as is compatible with the observation that the vertical slope of the front in the lower troposphere is toward the potentially coldest air. Hence there is established a zone in which the lid vanishes and the moist air becomes destabilized with respect to the growth of cumulus into cumulonimbus. Adjacent to the dry-line the lapse-rate is steep up to middle-levels in accordance with a westward increase of θ in the moist air, and there the initiation of the first large and showering cumulus can occur over favourable topographical sites. The most intense thunderstorms result from those cumulonimbus groups that propagate away from the dry-line but remain in the region of destabilization close to the boundary of the restrictive Mexican plume.

Lids appear to be associated with severe local storms other than those which occur over western Europe and Oklahoma. Thus the intense thunderstorms and line squalls over tropical Africa in summer are known to be associated with a restrictive inversion aloft produced by the advection of Saharan air over the moist southwesterly monsoon currents (as is shown on the

Niamey sounding, fig.14 ; see also Reiter, 1961, p.269). In addition, over India during the monsoon the prevalent thunderstorms appear to follow a significant rise in θ_w over the Arabian Sea, beneath a restrictive inversion produced by the advection aloft of a very warm adiabatic layer from Arabia (see Colon, 1964, p.194). The infrequency of severe local storms (suggesting that in general convective processes are accomplished adequately by the spectrum of cumulus sizes below large cumulonimbus) is consistent with the view that only when the normal process of convective energy transfer is impeded by a peculiar thermal stratification imposed by geographical features, can there be a storage of energy at low levels eventually released in a violent upheaval.

APPENDIX 1

Stratification accompanying cumulonimbus convection.

In this appendix several examples are given to support the statement made in the Introduction (p.12) that in cumulonimbus convection θ_s at the level of the anvil tops is close to the value of θ_w corresponding to the temperature at the observed level of the cloud bases, and that the ascent of air in cumulonimbus is much more nearly adiabatic than in cumulus.

Fig.87 shows the stratification of the troposphere over southern England on 2 July 1956. In the afternoon the screen-level θ_w inland reached 15.0 to 15.3°C when the value determined on the aircraft sounding at the cloud base was 14.1 to 14.5°C. On the composite sounding in the figure the adiabatic and cloud layers are readily recognised. The cumulus convection was widespread and vigorous ; the peak heights reached by the tallest towers before subsidence and evaporation were determined by the double-theodolite method to be between 3000 and 4500 m. The aircraft and radiosonde soundings show the top of the cumulus cloud layer at about 3500 m, with a lapse-rate through a deep layer above which it is only slightly less than the saturated adiabatic. Precipitation, indicated by the fibrillation of the decaying cloud towers or a radar echo, formed in those clouds whose towers reached or exceeded about 3500 m (fig.88). Some of the shower clouds grew rapidly to much higher levels, the peaks reaching 8400-8700 m before subsiding into anvils with flat tops at 7500 m.

The level of the observed anvil tops is very close to the equilibrium level determined by adiabatic ascent on the aerological diagram, indicating that in the larger cumulonimbus at least a proportion of the cloud air rose from the base with very little effective mixing.

On 25 June 1957 winds over England were light NW'ly and conditions inland nearly uniform. Radar observations showed

that numerous showers formed before noon when the cumulus tops reached above 3000 m. On this occasion the lapse-rate above the cumulus layer, indicated on the sounding (fig.89) at 3000 m, was more stable, and the cumulus towers reached peak heights of only 3500 m. At mid-day an experienced glider pilot used an updraught of 3 m/sec to reach 2900 m (temperature -8°C) in a cumulus whose top reached 3200 m. No precipitation or ice crystals were observed ; a layer of ice $\frac{1}{4}$ in thick formed on the glider. Soon afterwards very smooth strong lift was encountered below and in building cumulus on the flank of a shower cloud. Between 1300 and 2000 m the average speed of the updraught was $3\frac{1}{2}$ m/sec and from there to 3800 m was $7\frac{1}{2}$ m/sec. Ice accumulated to a thickness of 1 in, and between 3800 m and 4400 m small ice crystals entered the cockpit. The glider emerged from the side of a tower reaching 5000 m, which was approximately the peak height attained by the radar echoes from the afternoon showers. Although on this occasion the cumulonimbus were only small, with ill-defined anvil tops at about 4500 m, the experience of the glider pilot confirms the indication on the aerological diagram that their updraughts possessed a substantially greater buoyancy than the cumulus updraughts.

This difference is evident indirectly in observations of bigger clouds made inland over central Sweden on 10 July 1955. Ludlam and Saunders (1956), used double-theodolite measurements of cloud height to determine that on this day the peak heights reached by the cumulus towers which formed showers were close to 6800 m. The local mid-afternoon sounding (fig.90) was made near the shore of a large cool lake, and does not correctly indicate the somewhat higher values of θ and θ_w observed away from the shore. The local cumulus layer, with a top at about 4 km, and the deeper layer, up to 6300 m, modified by the more vigorous convection over the mountains (about 1 km high) 40 km or more upwind, are clearly represented. Cumulonimbus developed over the mountains and foothills during the afternoon, and towered to 11,000 m,

producing extensive anvils with tops at 9400 m. At this level θ_s on the sounding is again very close to the θ_s at the measured level of the cloud base. Fig.91, showing rates of rise of cloud towers as a function of height, is also taken from Ludlam and Saunders (1956), but has been slightly modified to separate the values more clearly according to whether or not the parent cloud was a shower cloud. It is quite clear that the cumulonimbus towers have about their maximum ascent rates, and probably therefore about their greatest buoyancy, at levels between 7 and 9 km which the cumulus towers could not even reach.

An example of cumulonimbus convection in a more unstable stratification, but still with little wind shear, is provided by observations of thunderstorms inland over southern England on the afternoon of 18 June 1957, which have been described and discussed by Ludlam (1962). The radiosonde sounding made at Crawley about 3 hours before the thunderstorms developed in the afternoon (fig.92) shows the local cumulus layer extending up to 3700 m, but the lapse-rate steepens again above this layer in air which had been modified by deeper cumulus convection over Belgium on the previous afternoon. The top of the higher cumulus layer is indicated at 6000 m, and in mid-afternoon the cumulus formed over southern England quickly re-occupied this layer. Showers formed in the largest clouds, whose towers reached 6700 m before subsiding. Within an hour of the first shower formation cumulonimbus developed with tops approaching the tropopause, and in the period of maximum intensity between two and three hours later isolated peaks reached 13,200 m above moderately extensive anvil clouds with tops at 10,000 m. On the sounding θ_s at this level is appreciably less than the value appropriate to the cloud bases. It appears that on this occasion the lack of wind shear prevented the efficient organisation of the cumulonimbus convection, for the clouds propagated impulsively and not continuously, and the ascent of the cloud air was evidently accompanied by a substantial mixing which permitted the release

of only a fraction of the potential energy apparently available. The thunderstorms which occurred were only moderately severe, whereas on 5 September 1958, when the instability as conventionally assessed was about the same, there was a more favourable wind shear and the efficient energy release resulted in the violent Horsham storm discussed in the text.

APPENDIX 2

Penman (1948) has simplified the computation of L over reasonably moist terrain by use of the Bowen ratio and the surface energy balance. Assuming that the eddy diffusion is responsible for the transfer upwards of both heat and water vapour, the Bowen ratio can be expressed as :

$$H/L = 0.64(K_H/K_W)(p/1000)(T_s - T_a)/(e_s - e_d) = 0.64(T_s - T_a)/(e_s - e_d)$$

where, (K_H/K_W) is the ratio of the eddy diffusion coefficients of heat and water vapour. Priestley has shown that it is about 1 under neutral conditions

p is the surface pressure in mb (taken here as 1000),

T_s and T_a are the surface and screen-level temperatures in $^{\circ}\text{C}$,

e_s and e_d are the surface and screen-level vapour pressures in mb.

Three assumptions are necessary to solve for L in terms of usable quantities. The first is

$$L \text{ (calculated)} \equiv (e_s - e_d)f(u) \quad \text{where the definition}$$

$$L_a \equiv (e_s - e_a)f(u) \quad \text{and}$$

e_a is the saturation vapour pressure of the atmosphere at screen level : $e_a = e(T_a)$;

$f(u)$ is a function of wind speed at screen level.

The second assumption is that

$$\partial T / \partial e \text{ (saturated)} \equiv (T_s - T_a)/(e_s - e_a) = \text{constant at } T_a$$

APPENDIX 2 (continued)

The third assumption is that

$$e_s = e(T_s)$$

Assumptions two and three will be invalid over arid terrain and hence the following equations apply only to cultivated regions.

Combining the surface energy balance and the Bowen ratio

$$L_{(calc)} = \frac{R_n + 0.64 \left(\frac{p_s c_s \sqrt{x}}{p c_p \sqrt{k}} + 1 \right) \frac{\partial T}{\partial e} L_a}{1 + 0.64 \left(\frac{p_s c_s \sqrt{x}}{p c_p \sqrt{k}} + 1 \right) \frac{\partial T}{\partial e}}$$

For regions where fields, crops and forests are the natural vegetation and evaporation is largely unimpeded by lack of moisture, the above equation reduces to

$$L_{(calc)} = 0.8 R_n + 0.3 (e_a - e_d) f(u)$$

Experiments have shown that the ratio $L/L_{(calc)}$ is about 1.0 over open water but will be about 0.8 over a wide variety of cultivated soil types in summer. Use of the formula, adapted from Penman, $f(u) = 16.B$, where B is the Beaufort wind number, and the relation $(e_a - e_d) = 1.7(T_a - T_d)$ yields

$$L = 0.6 R_n + 7 \overline{(T_a - T_d)B}$$

where the bar on the right hand term refers to a mean product averaged over a solar day.

On a typical cloud-free mid-summer day the net radiation R_n at middle latitudes will be about 350 lys and the energy partition can be expressed

$$\begin{aligned} L &= 220 + 7 \overline{(T_a - T_d)B} \quad \text{ly/solar day} \\ H &= 100 - 5 \overline{(T_a - T_d)B} \end{aligned}$$

Obviously, then, L must be considerably greater than H over the normal range of temperature-dewpoint excess and wind speeds.

APPENDIX 3

The isentropic charts presented for the five cases of severe storms illustrate a model of baroclinic flow which can be related to certain observations and notions concerning fronts and cloudiness. Fig.93 is a schematic diagram of relative streamlines on an isentropic surface through a typical large-scale baroclinic wave disturbance. In a sense it is a horizontal aspect of the schematically drawn baroclinic cell shown in fig.1. The main driving branch in this quasi-horizontal cell is an ascending and anticyclonically curved stream leading from the layer of small scale convection in the south into a narrow confluence zone ahead of the trough axis where frontal cloud is formed. The other driving branch, that of cold air descending, originates behind the trough and splits into two streams - one continuing to turn anticyclonically in descent behind the front and the other flowing on a cyclonic path around the trough and ascending somewhat. The front is shown as an air mass (but not necessarily a temperature) discontinuity which can be represented by a surface of 'limiting streamlines' between the two main branches.

In considering a family of isentropic analyses the model demonstrates that the ascending branch provides the main injection of potentially warm air into the upper troposphere. Successive streams in the vertical become stacked one upon the other in the confluence zone with the warmest plume arriving above the others. Consistent with isentropic surfaces sloping upward toward the cold air the greatest total ascent occurs nearest the front and hence cloud formation and vertical motion are most pronounced in that part of the system. The front itself may be slanted toward the cold air and the tropopause ahead of it in the ascending warm branch will be high. The descending branch also acquires heat and moisture near the ground but cloud is mainly cumuliform. In the ascending part of the cold stream west of the front the cumulus may grow to considerable height and even reach cumulo-

APPENDIX 3 (continued)

nimbus proportions, although cloud tops will be restricted by the low tropopause.

The anticyclonic and ascending stream involves a thermally direct circulation, and the rise of potentially warm air into the confluence zone is accompanied by the conversion of potential into kinetic energy, manifested by a horizontal acceleration, particularly nearest the front and in the warmest plumes. It was often possible on the soundings to locate a wind maximum in or slightly above the air streams which originated in a particularly warm layer of small-scale convection. In the Horsham case, for instance, on both the Crawley and Chateauroux soundings (figs. 19, 22) wind speed maxima near the 650, 450 and 320 mb levels correspond to streamlines originating in the convective layers over Spain, the Sahara, and in the cloud layer over the sub-Sahara, respectively. Even more striking is the strong wind speed maximum at 700 mb in the Spanish adiabatic layer aloft over Camborne (fig.43) on 29 August, 1961. Over the plains of the United States a wind maximum is consistently observed at the top of the ascending Mexican adiabatic layer, particularly near the limiting streamline (see soundings, figs. 51, 65, 67, etc.). As Riehl and Fultz (1957, p.229) have implied in idealized tank experiments of baroclinic motion, the axis of the jet stream is located in the confluence zone along the limiting streamline, as the result of direct ascent in the warm streams of fluid.

It is not surprising, therefore, that the frontal cloud is often sharply delineated along the edge of the anticyclonic baroclinic cell, as noted by Reiter (1961, pp.360-364) and that the front and jet stream axis tend to coincide with the limiting streamline, which signifies a discontinuity in the weather. Boucher et al (1963, p.158) have presented a ~~Tyros~~^{Tyros} satellite picture showing the cloud distribution associated with a typical mature middle-latitude cyclone. The photograph strongly suggests

APPENDIX 3 (continued)

that the sharply-defined western edge of a long belt of frontal cloud, resembling that in cases presented here (see fig.13, for example), corresponds to a limiting streamline. To the west of the limiting streamline lies a dry tongue, described by Riehl (1962, p.59), which corresponds to air free from cloud which has recently descended.

While the migrating eddies are manifested by the continuous intrusion of potentially warm air into middle latitudes, the geography imposes certain fixed sources of air with a high θ_w highly favourable for ascent in quasi-horizontal slant convection. Indeed, the primary coupling between the circulations of high and low latitudes occurs in the tilted troughs which form an interruption of the sub-tropical belt of easterlies (see Arakawa, 1953). These troughs are to be found as semi-permanent features of the weather off the west coast of north Africa in summer, and elsewhere on the west side of ocean areas. Coupling between the African trough (the counterpart of the Saharan anticyclone) and deep migratory troughs approaching Europe is illustrated in the three European cases.

REFERENCES.

- Arakawa, H. 1953 Tilted-trough model as an interaction between low and high latitude disturbances. J. Meteor., 10, p.64.
- Ball, F. K. 1960 Effect of heating on inversions. Quart. J. R. Met. Soc., 86, p.483
- Boucher, R. J. et.al. 1963 Synoptic interpretations of cloud vortex patterns as observed by meteorological satellites. Final Report, Aracon Geophysics Co., Nov. 1963. Contract No. Cwb-10630. U.S. Weather Bureau reprint.
- Browning, K.A. and Ludlam, F. H. 1962 Airflow in convective storms. Quart. J. R. Met. Soc., 88, p.117
- Browning, K.A. and Donaldson, R. J. 1963 Airflow and structure of a tornadic storm. J. Atmos. Sci., 20, p.533
- Bunker, A. et.al. 1949 Vertical distribution of temperature and humidity in the Caribbean Sea. Woods Hole Papers in Physical Oceanography and Meteorology, XI, No.1. M.I.T. and Woods Hole Oceanographic Institution, Massachusetts.
- Colon, J. 1964 On interactions between the southwest monsoon current and the sea surface over the Arabian Sea. Indian J. of Met. and Geophys., 15, p.183.
- Craddock, J.M. 1949 The development of cumulus cloud. Quart. J. R. Met. Soc., 75, p.147
- Cunningham, R., Glass, M. and Carlson, T.N. 1964 Properties of active cumulus clouds determined from co-ordinated ground photography and aircraft penetrations. Unpublished manuscript. Presented at National Conference on the Physics and Dynamics of Clouds. March 1964. Chicago.
- Dhonneur, M. 1957 Etude climatologique des routes aériennes aboutissant a Fort Lamy. Monographs de la Météorologie Nationale, No.2. Paris.
- Douglas, C.K.M. and Harding, J. 1946 The thunderstorm of the night of July 14-15, 1945. Quart. J. R. Met. Soc., 72, p.323

- Dyer, A. J. 1961 Measurements of evaporation and heat transfer in the lower atmosphere by an automatic eddy-correlation technique. Quart. J. R. Met. Soc., 87, p.401
- Fawbush, E.T. and Miller, R.C. 1954 The types of air masses in which North American tornadoes form. Bull. Amer. Met. Soc., 35, p.154
- Fujita, T. 1958 Structure and movement of a dry front. Bull. Amer. Met. Soc., 39, p.574
- 1959 Precipitation and cold air production in mesoscale thunderstorm systems. J. Met., 16, p.454
- Halstead, M.H. 1954 Publications in climatology. John Hopkins University, 7, p.353
- Hamilton, R.A. and Archbold, J.W. 1945 Meteorology of Nigeria and adjacent territory. Quart. J. R. Met. Soc., 71, p.231
- Hawson, C.L. and Caton, P.G.F. 1961 A synoptic method for the international comparison of geopotential observations. Met. Mag., 90, p.336
- Howell, W.C. 1960 Cloud seeding in the American tropics. Geophys. Monographs No.5. Amer. Geophys. Union. Washington, D.C.
- James, D.G. 1953 Fluctuations of temperature below cumulus clouds. Quart. J. R. Met. Soc. 79, p.425
- Jones, R.F. 1960 Size distribution of ice crystals in cumulonimbus clouds. Quart. J. R. Met. Soc., 86, p.187
- Lettau, H. 1951 Theory of surface temperature and heat transfer oscillations near a level ground surface. Trans. Amer. Geophys. Union, 32, p.189
- Ludlam, F.H. 1962 A case study of thunderstorms on an occasion of small wind shear. Technical (Scientific) Note No.11. Imperial Coll. Dept. of Meteorology.
- 1963 Severe local storms, a review. Meteor. Monographs, 5. No.27., p.1. Amer. Met. Soc.

- Ludlam, F.H. and Macklin, W.C. 1959 Some aspects of a severe storm in S.E. England. *Nubila*, II, No.1, p.38
- Ludlam, F.H. and Macklin, W.C. 1960 The Horsham hailstorm of 5 September 1958. *Met. Mag.*, 89, p.38
- Ludlam, F.H. and Saunders, P.M. 1956 Shower formation in large cumulus. *Tellus*, 8, p.424
- Malkus, J.S. 1954 An example of orographic-convection cell in the trade-wind region. Woods Hole unpublished manuscript. Tech. Report No.30. Ref.54-43. Contract N 60 nr-27702. Woods Hole Oceanographic Institution, Mass.
- 1962 Large-scale Interactions. *The Sea*, 1, p.208. Interscience Publ., N.Y.
- McGuire, E.L. 1962 The vertical structure of three dry lines as revealed by aircraft traverses. NSSP, Report No.7 U.S. Weather Bureau.
- Miller, R.C. 1959 Tornado-producing synoptic patterns. *Bull. Amer. Met. Soc.*, 40, p.465
- Moller, F. 1951 Long wave radiation. *Compendium of Meteorology*, p.35. Amer. Met. Soc.
- Monteith, J.L. 1962 Attenuation of solar radiation : a climatological summary. *Quart. J. R. Met. Soc.*, 87, p.159
- Monteith, J.L. and Sceicz, G. 1961 The radiation balance of bare soil and vegetation. *Quart. J. R. Met. Soc.*, 87, p.159
- National Severe Storms Project (NSSP), Staff Members 1963 Environmental and thunderstorm structures as shown by National Severe Storms Project observations in Spring 1960 and 1961. *Mon. Wea. Rev.*, 91, p.271
- Newton, C.W. 1963 Dynamics of severe convective storms. *Meteor. Monographs*, 5, No.27, p.33. Amer. Met. Soc.
- Pasquill, F. 1949 Some estimates of the amount and diurnal variation of evaporation from a clayland pasture in fair spring weather. *Quart. J. R. Met. Soc.*, 75 p.249
- Penman, H.L. 1948 Natural evaporation from open water, bare soil, and grass. *Proc. Roy. Soc. London, A*, 193, p.120

- Penman, H.L. 1956 Evaporation : an introductory survey. Netherlands J. of Agricultural Sci., 4, p.9
- Petterssen, S. et.al. 1946 Convection in theory and practice. Meteorological Office, Air Ministry Synoptic Division's Technical Memorandum. No.102. MRP 242
- Priestley, C.H.B. 1959 Turbulent transfer in the lower atmosphere. , Univ. of Chicago Press.
- Pruitt, W.O. and 1963 Investigations of energy and mass
Aston, M.J. transfer near the ground including influences of the soil-plant atmosphere. Final Report June 1963. Univ. of California, Davis. Depts. of Agricultural Engineering and Irrigation.
- Reiter, E.R. 1963 The structure of jet streams. Univ. of Chicago Press.
- Riehl, H. 1962 Jet streams in the atmosphere. Tech. Report No.32. Colorado State University. Dept. of Atmospheric Physics.
- Riehl, H. and 1957 Jet streams and long waves in a
Fultz, D. steady rotating dishpan experiment. Quart. J. R. Met. Soc., 83, p.215
- Saunders, P.M. 1957 The thermodynamics of saturated air : a contribution to the classical theory. Quart. J. R. Met. Soc., 83, p.342
- Thorkelsson, T. 1946 Cloud and shower. Quart. J. R. Met. Soc., 72, p.332
- Warner, J. 1963 Observations relating to theoretical models of a thermal. J. Atm. Sci., 20, p.546

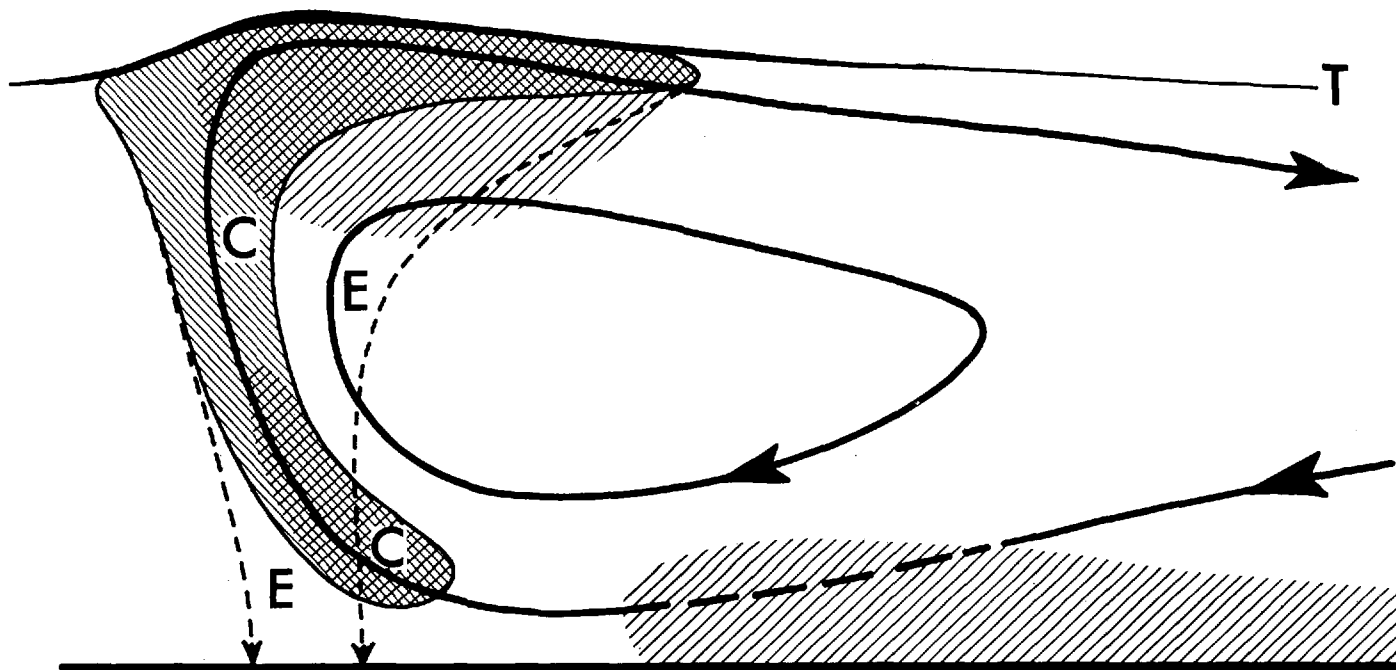


Fig. 1. Schematic representation of tropospheric convection in vertical section. The hatched area at the surface represents the layer of small-scale convection, within which subsiding air has its potential temperature raised. Air from this layer enters a cumulonimbus or large-scale circulation and ascends into the upper troposphere. Condensation (C) of water vapour occurs, and small-scale convection may develop again in the cloud system where there is cross-hatching. Precipitation falls from the cloud (pecked lines) and there is some evaporation (E) outside it. The descending branch of the large-scale circulation is left open on the right to indicate that it continues in several motion systems, accompanied by a slow radiative loss of potential temperature. T marks the tropopause.

Fig.2

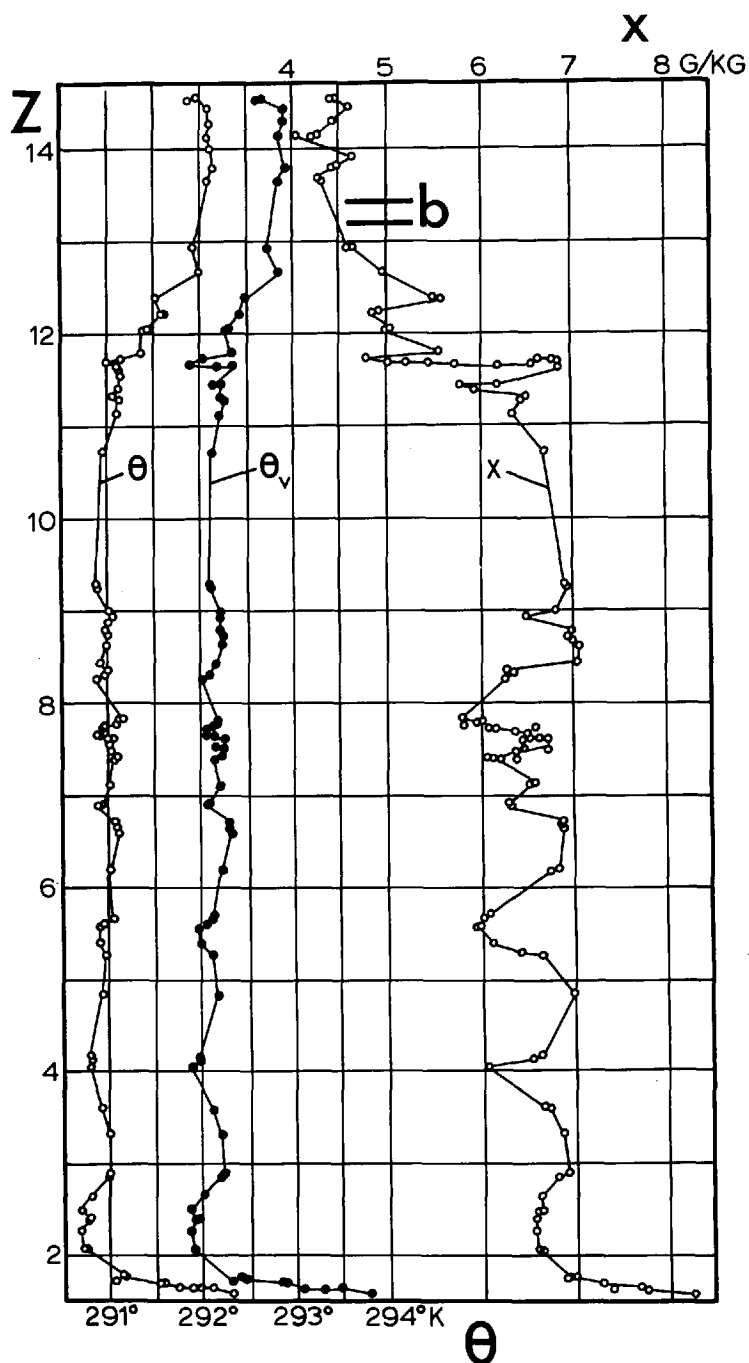


Fig.2. Sounding made by Miss Woodward over southern England from 1140 to 1153 G.M.T., 18 July 1961. Potential temperature θ , virtual potential temperature θ_v , and mixing-ratio x are shown as a function of height Z above sea level, in metres. The lowest level of observation is at about screen-level. The cumulus bases were observed between the levels marked b .

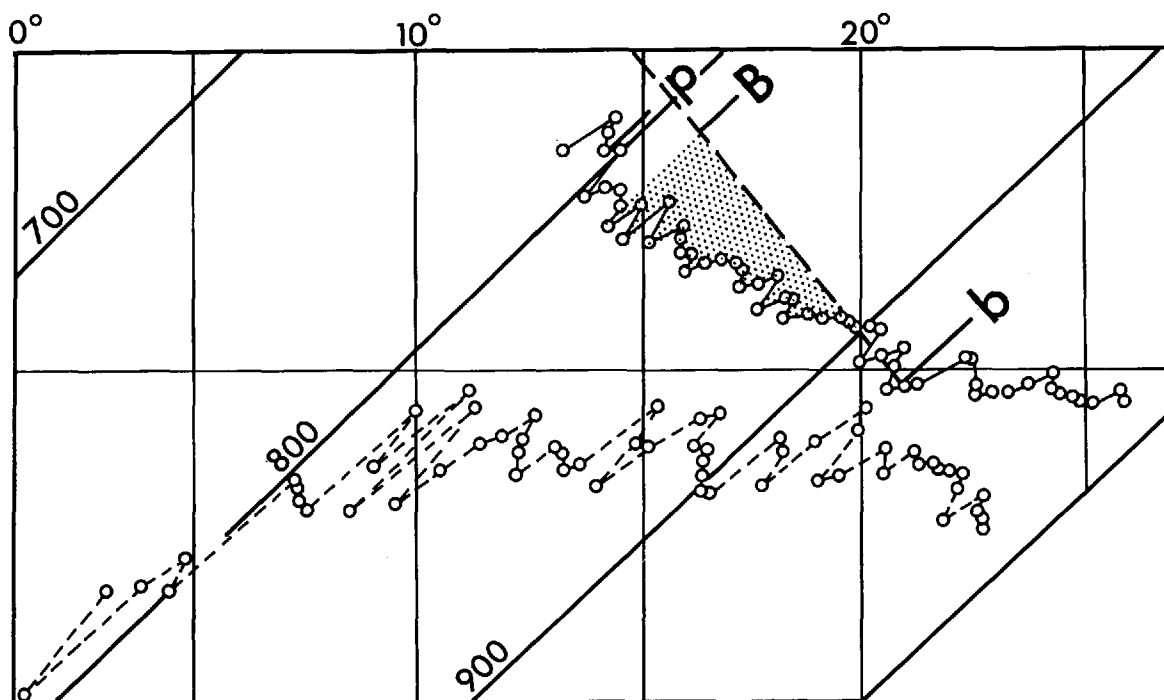
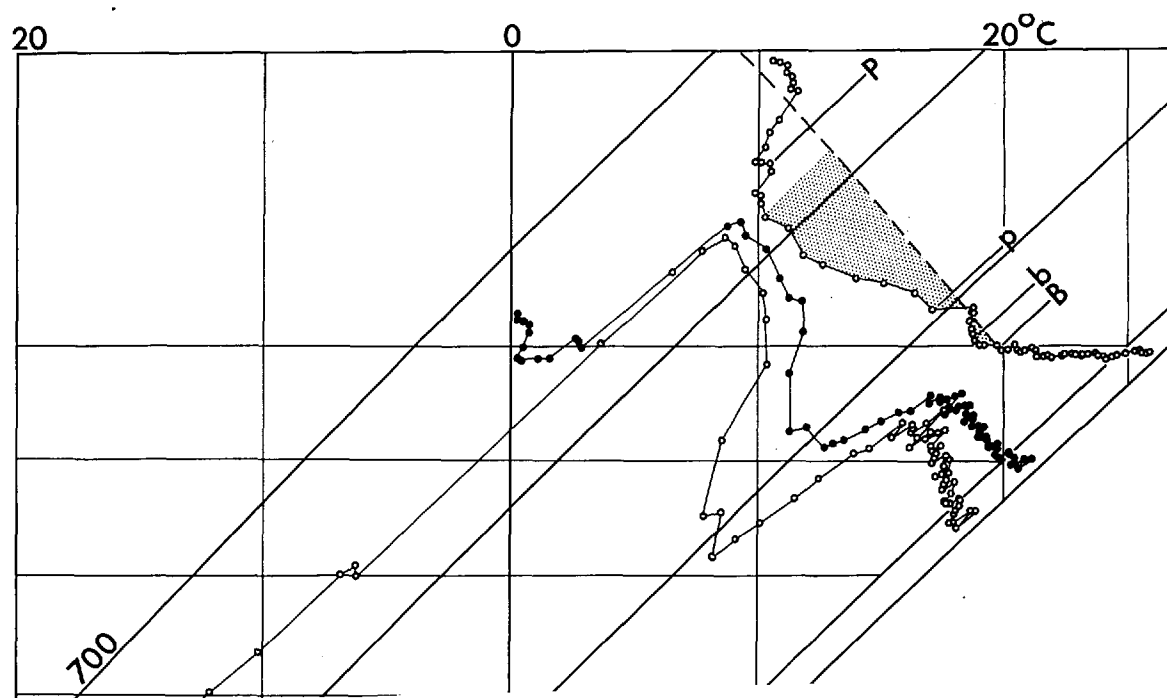


Fig. 3. Sounding from Malkus (1954), made inland over Puerto Rico near noon, 25 June 1952: temperature and dew-point as a function of height, represented on a tephigram (the pecked line is a saturated adiabatic). The observed level of cumulus bases is marked *b*, and that of their tops by *p*; *B* marks the top of the stippled layer in which the mean lapse-rate is greater than the saturated adiabatic (the cloud layer). The isobars are labelled in mb.



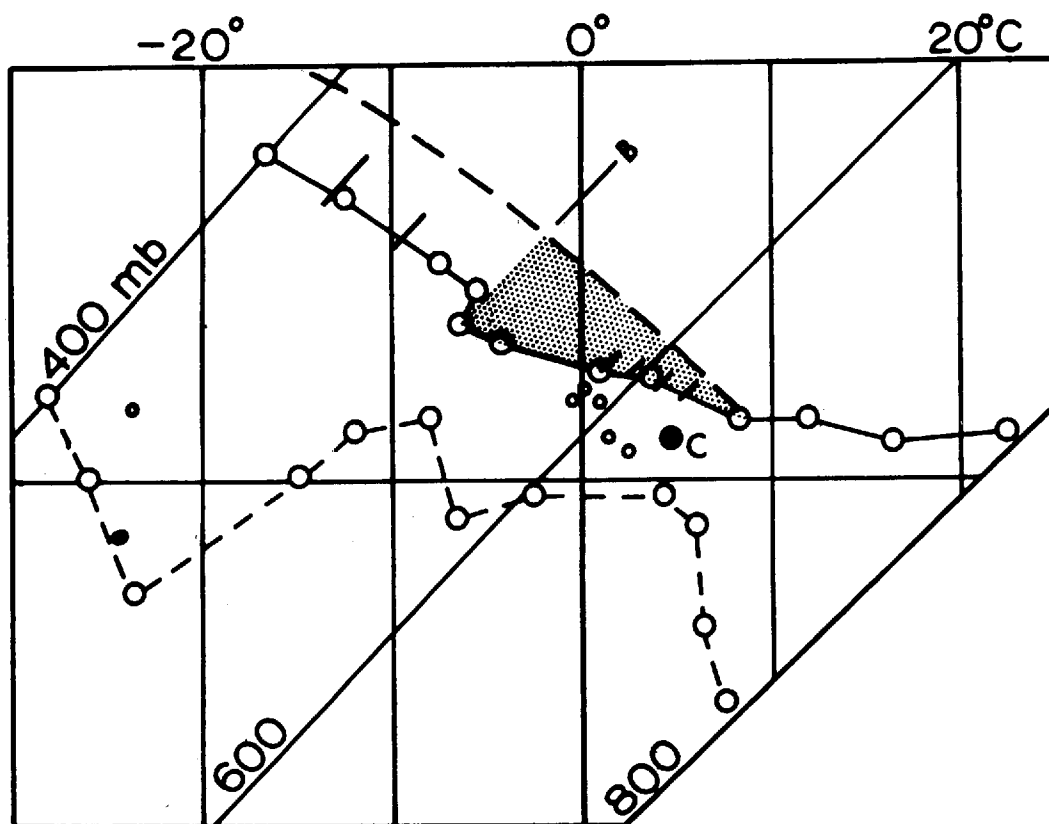


Fig. 5. Radiosonde sounding at Flagstaff, Arizona about noon, 1 August, 1961. The pecked line is the saturated adiabatic through the point on the sounding corresponding to the level of cumulus base observed by an aircraft over mountains about 15 miles north of the radiosonde station; the point corresponding to the convective condensation level of the screen-level air at the station is marked C, and the top of the cloud layer B. Dew-points measured by the aircraft outside clouds are marked by small circles; the ranges of temperature measured inside clouds at several levels are shown by lines parallel to the isobars and intersecting the radiosonde sounding: those at the two uppermost levels were associated with shower clouds.

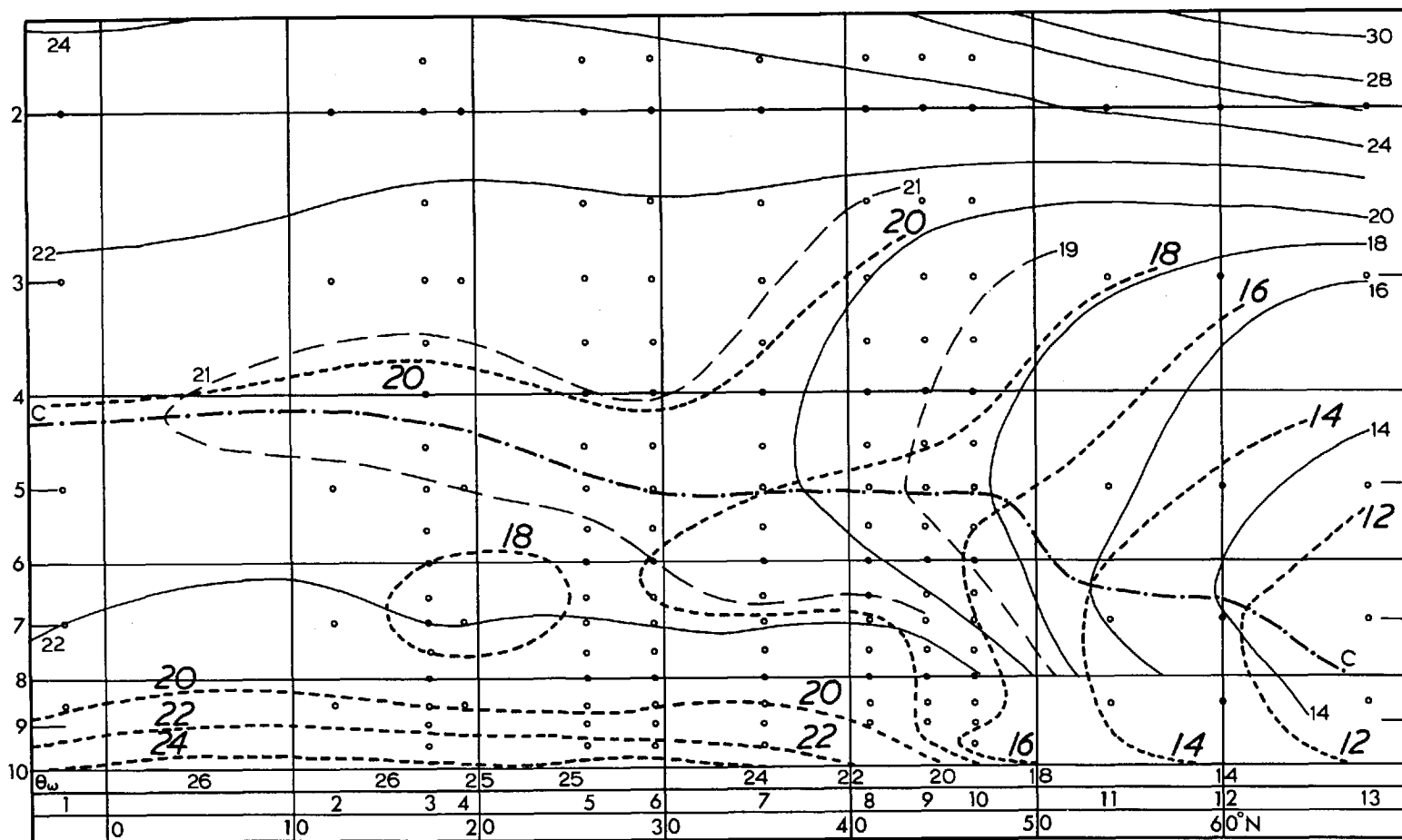


Fig.6. Mean summer vertical section over land in the Americas, approximately along the 90th meridian, July 1959. Thin solid lines show the distribution of Θ_s at intervals of 2°C ; thicker pecked lines the distribution of Θ_w (labelled in larger, italic figures). The dot-dash line marked C shows the 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: Guayaquil, 84129; 2: San Andres, 80001; 3: Swan Island, 78501; 4: Grand Cayman, 78383; 5: Brownsville, 72250; 6: San Antonio, 72253; 7: Oklahoma City, 72353; 8: North Platte, 72582; 9: Rapid City, 72662; 10: Bismarck, 72764; 11: The Pas, 72667; 12: Fort Smith, 72934; 13: Coppermine, 72938).

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 $^\circ\text{C}$.

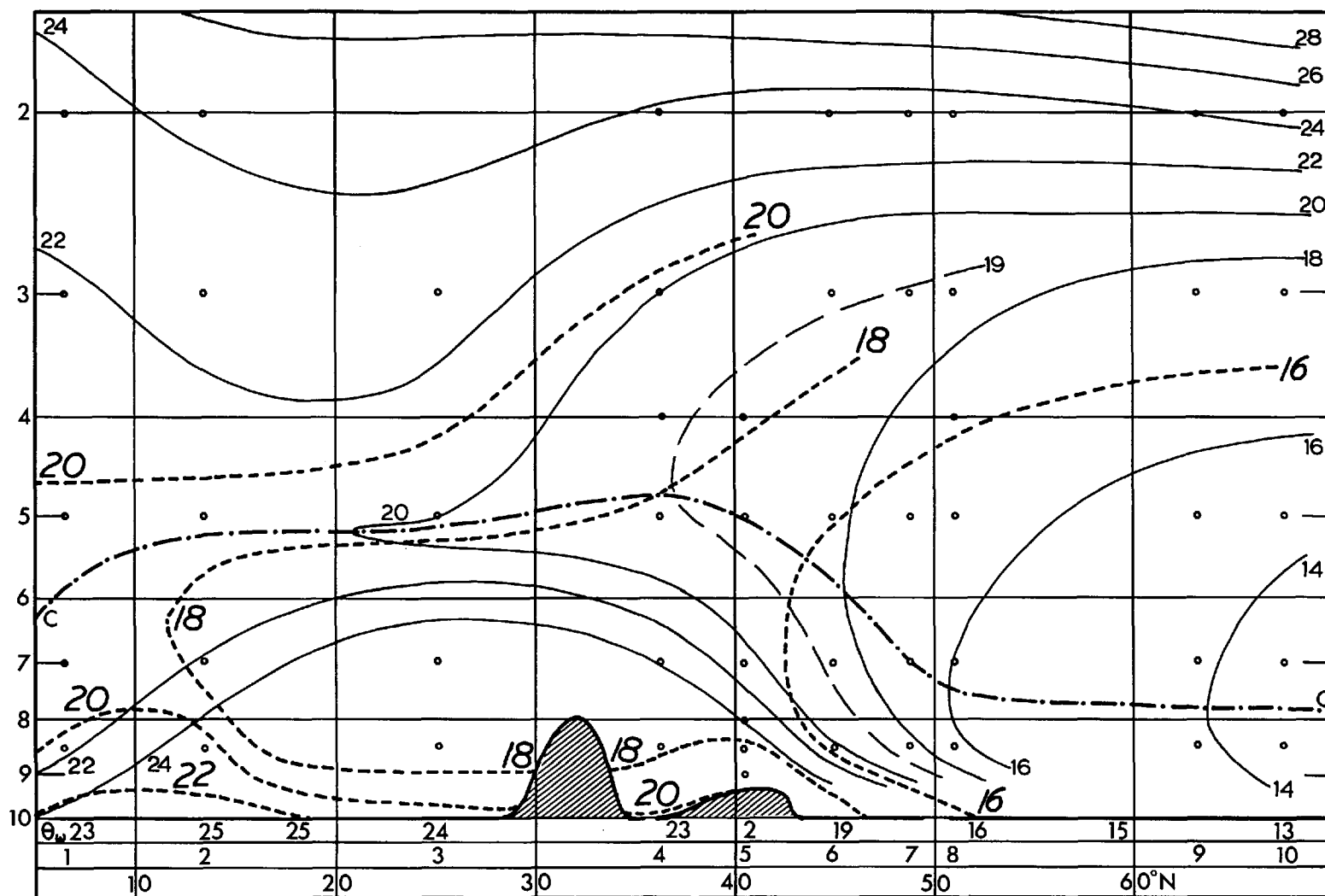


Fig. 7. Mean summer vertical section over land approximately along the Greenwich meridian, July 1959. For general explanation see legend to Fig. 6. (Soundings used from 1: Lagos, 65201; 2: Niamey, 61052; 3: Fort Trinquet, 61401; 4: Gibraltar, 08495; 5: Zaragoza, 08159; 6: Bordeaux, 07510; 7: Trappes, 07145; 8: Crawley, 03774; 9: Ostersund, 02062; 10: Sodankyla, 02836.) The Atlas mountains, on the northern edge of the Sahara, and Spain are shown schematically.

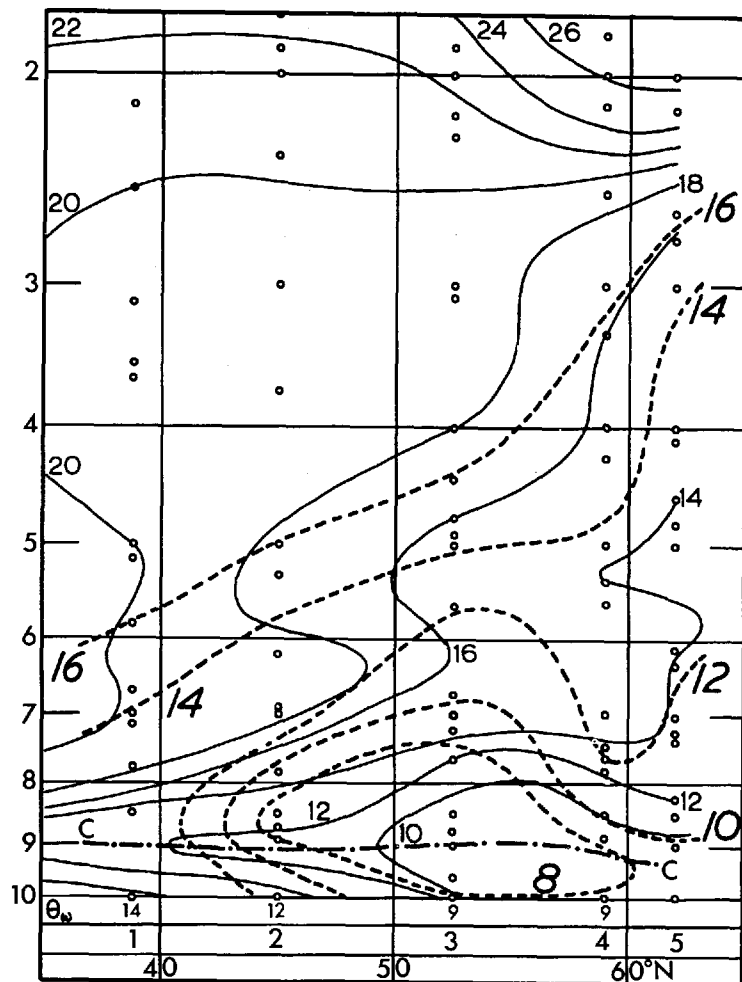


Fig.8. Vertical section over ocean, approximately along 20th meridian, in rear of a large-scale trough, 9 July 1959. For general explanation see legend to Fig.6. (Soundings used from 1: Lagens, 08509; 2: Ship K; 3: Ship J; 4: Ship I; 5: Ship A.)

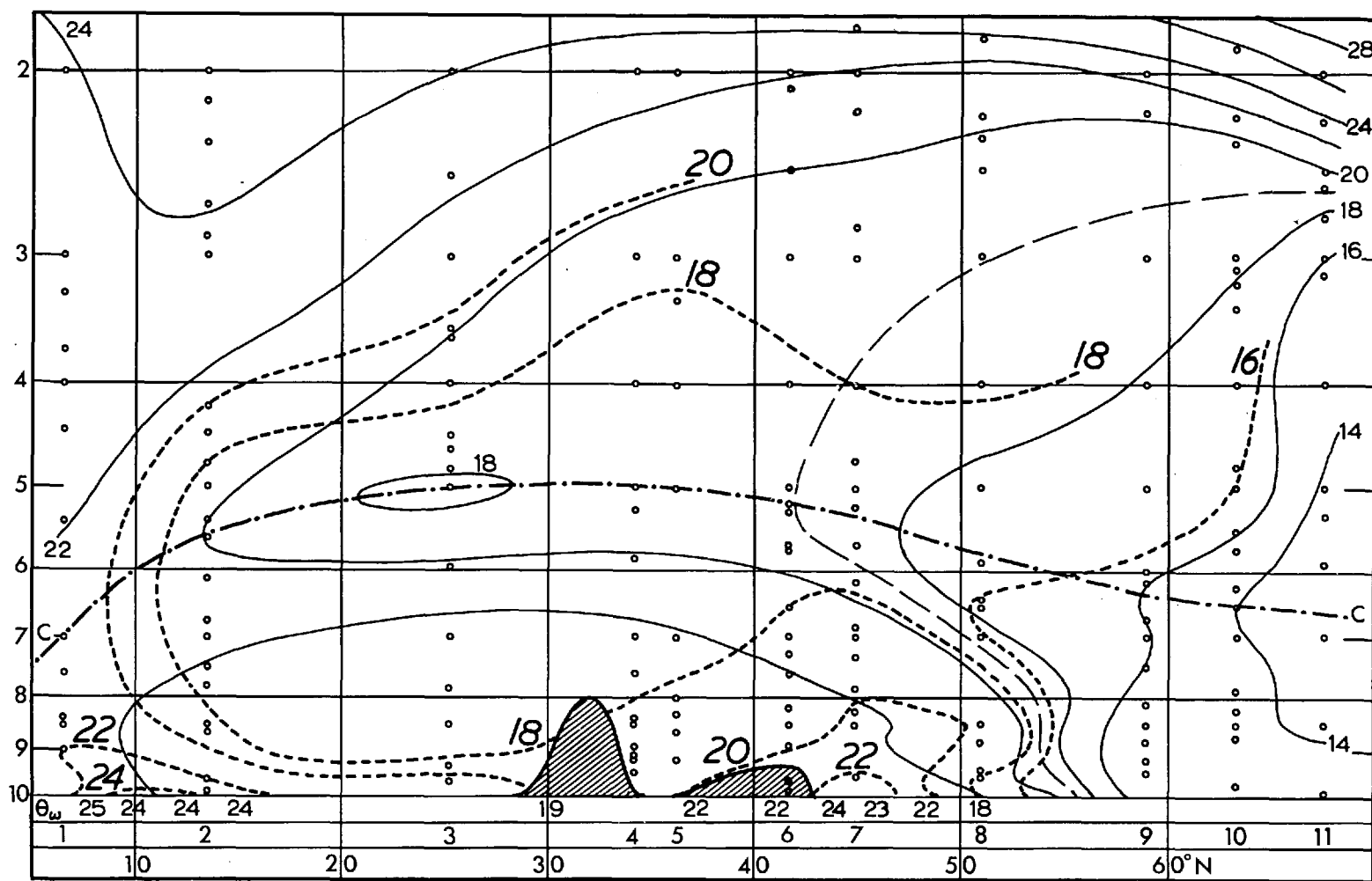


Fig.9. Vertical section over land, approximately along Greenwich meridian, in advance of a large-scale trough, 9 July 1959. For general explanation see legend to Fig.6. (Soundings used from 1: Lagos, 65201; 2: Niamey, 61052; 3: Fort Trinquet, 61401; 4: Kenitra, 60119; 5: Gibraltar, 08495; 6: Zaragoza,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

Fig.10

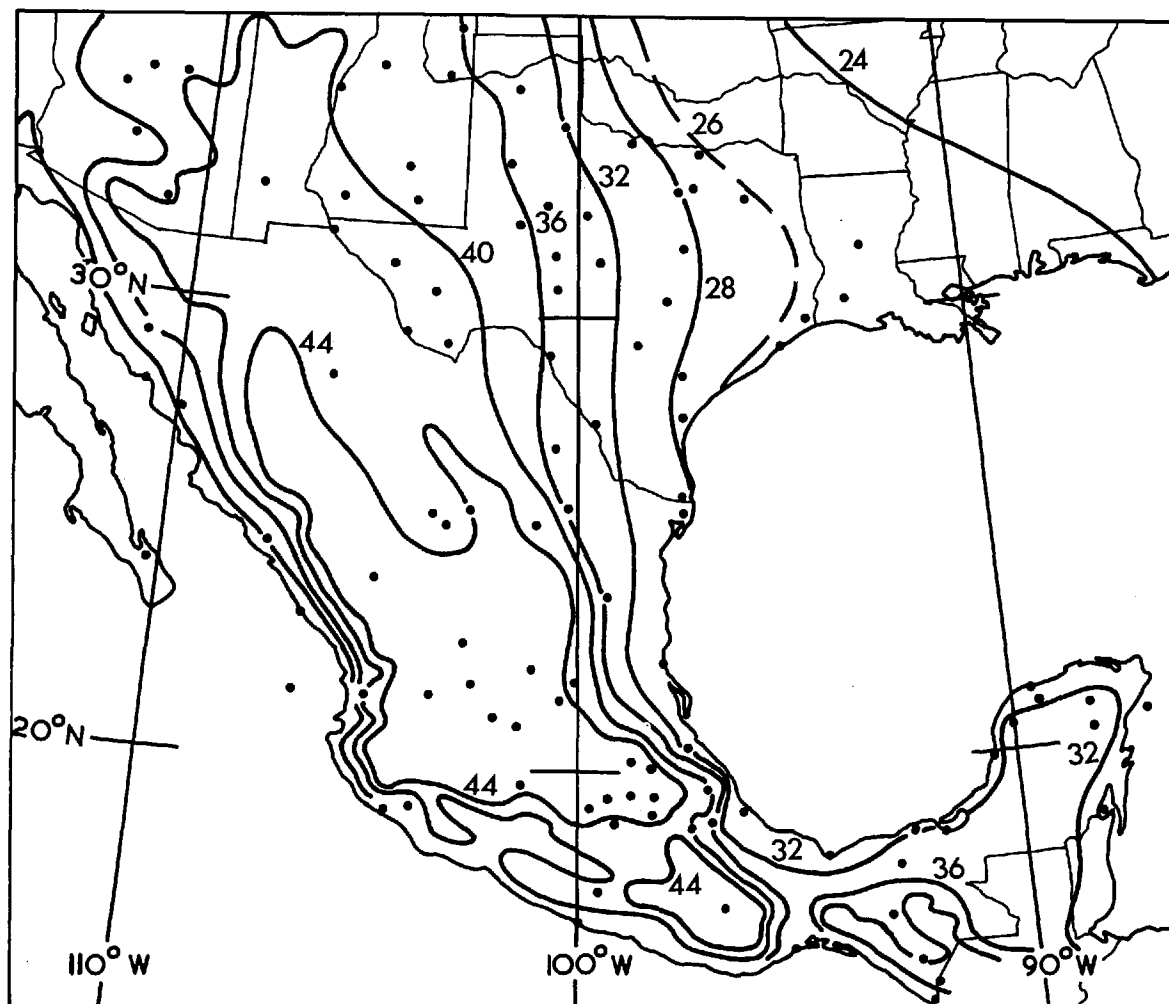


Fig.10. Distribution of mean maximum screen-level potential temperature (in $^{\circ}\text{C}$) over Mexico and adjoining territory during late April and early May.

(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.)

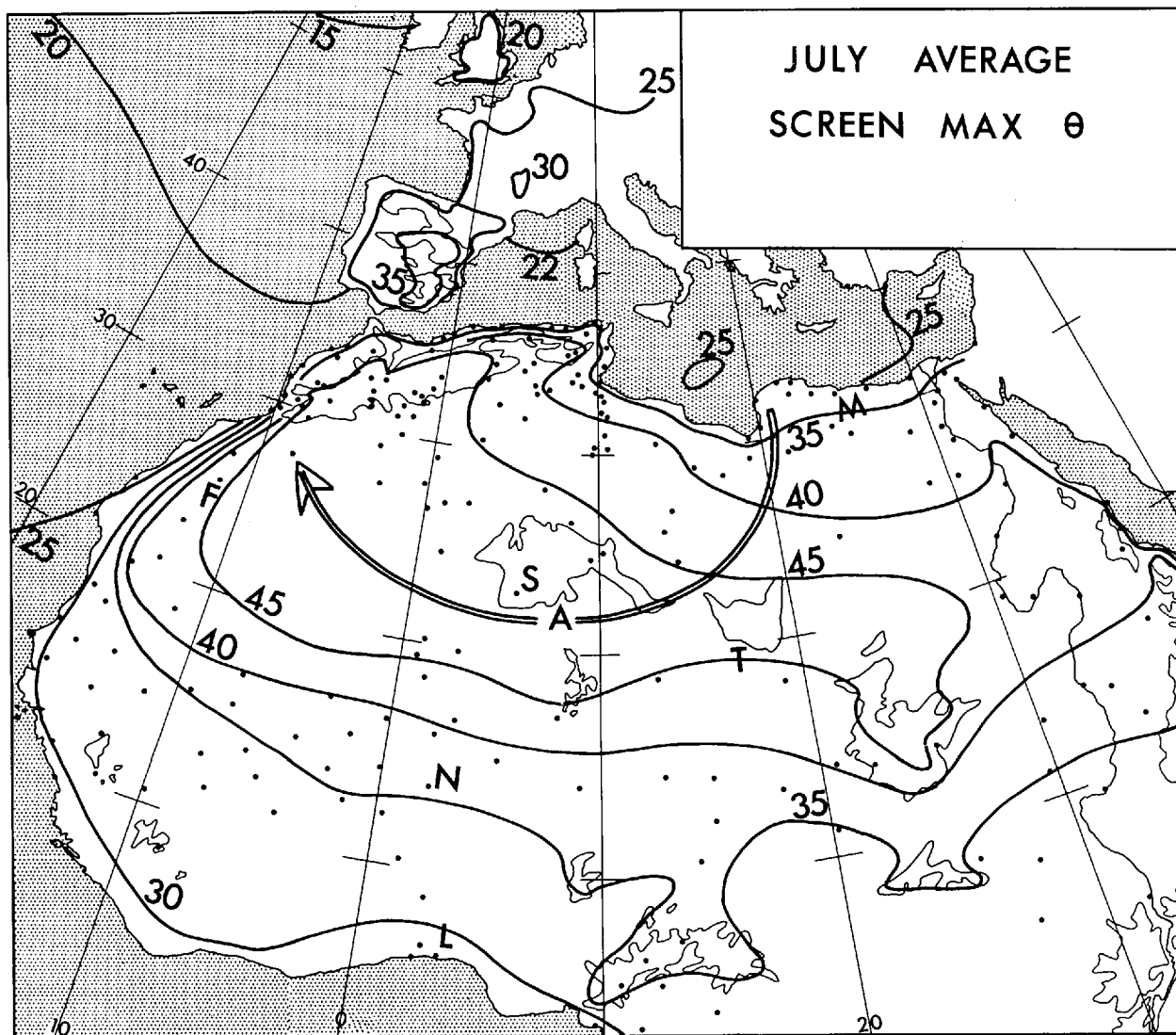


Fig. 11. Distribution of mean maximum screen-level potential temperature (in °C) over North Africa and western Europe during summer. (Values obtained from 30-year climatological means for July. Dots show the location of reporting stations, amongst which some mentioned in the text are marked as follows: M: Tobruk; S: Tamanrasset; N: Niamey; L: Lagos; F: Fort Trinquet. The thin lines are height contours for 3000 ft, and the Ahaggar and Tibesti mountains are marked by the letters A and T).

The diagram includes a streamline over the Sahara for the mean flow in the layer between 850 and 700 mb during July 1959.

Fig. 12

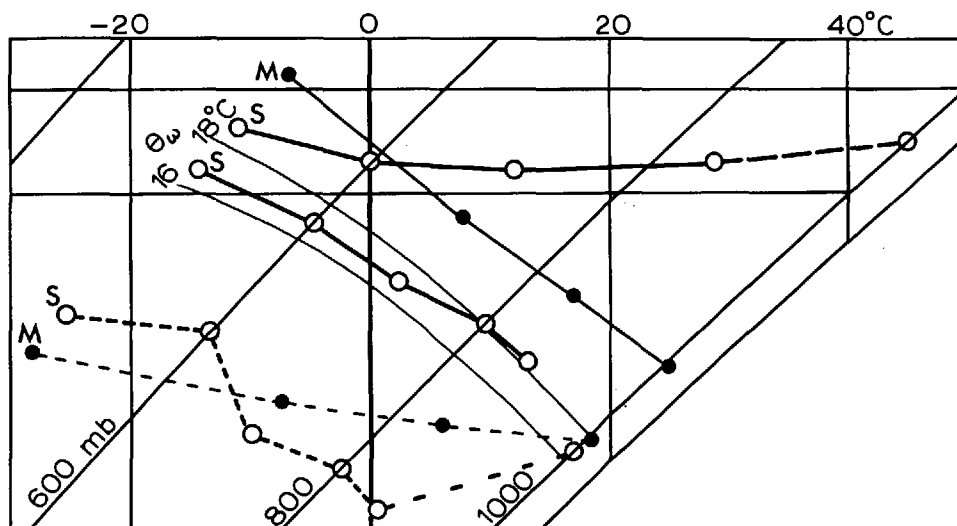


Fig. 12. Tephigram showing mean soundings for July (1959) at Tamanrasset (including wet-bulb temperatures, and labelled S) in the central Sahara, and upwind on the Mediterranean coast at Tobruk (full circles, labelled M; see also Fig. 11).

Fig.13

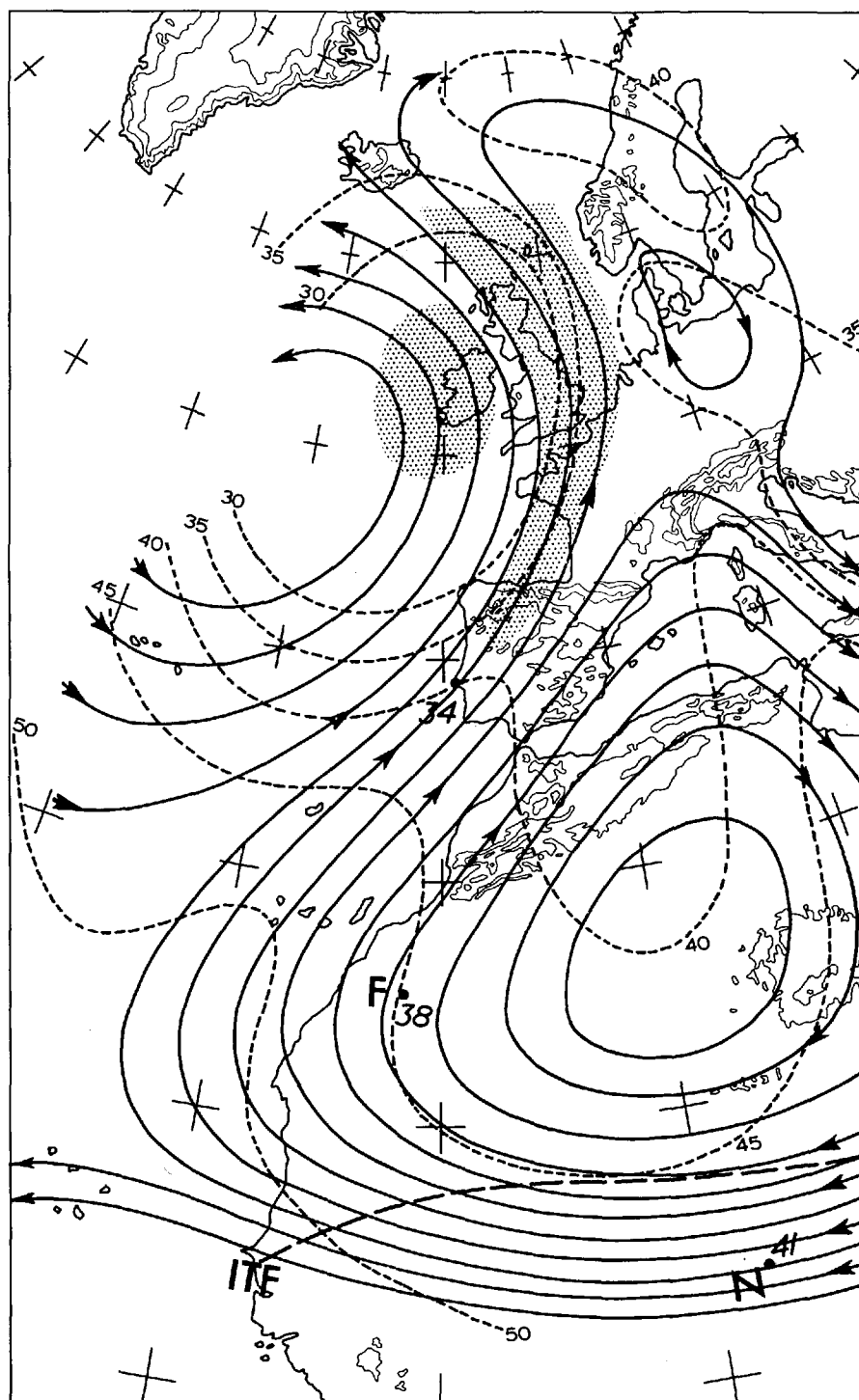


Fig.13. Isentropic relative-flow chart for $\Theta = 52^{\circ}\text{C}$, 1200 G.M.T., 5 September 1958, containing streamlines (solid) and isobars (pecked lines labelled in tens of mb). The pecked line marked ITF is the surface position of the intertropical front, and the letters N and F mark the positions of Niamey and Fort Trinquet. Italic figures give the adiabatic lifting condensation level (in tens of mb) at these places and Lisbon, and the stripling shows areas of extensive high cloud. The height contours for 3000 ft are drawn as thin lines.

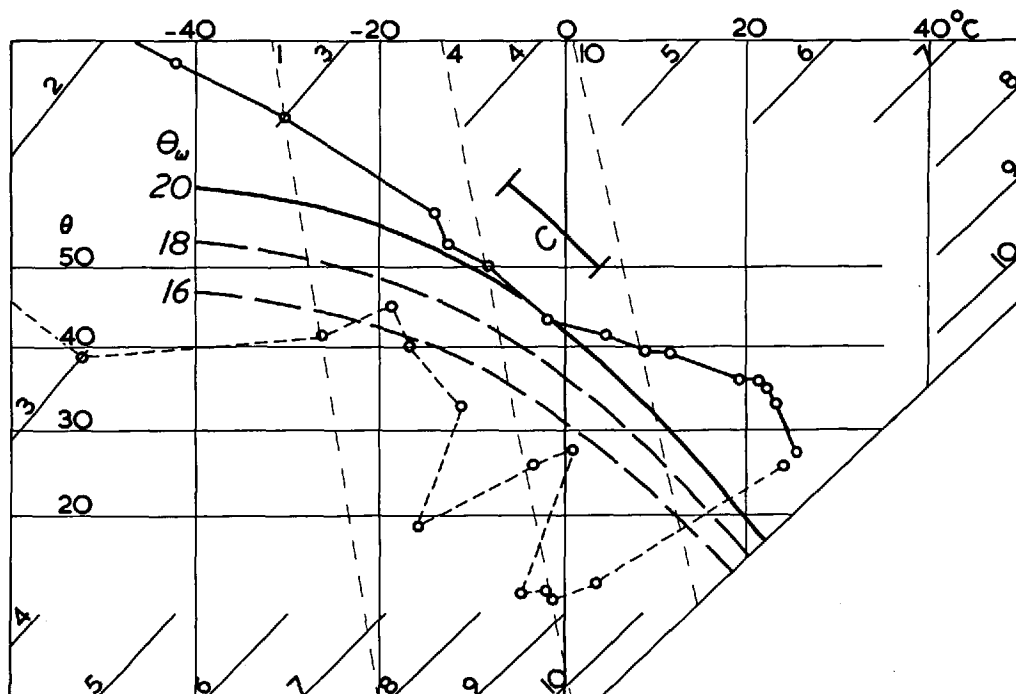


Fig. 14. Niamey (61052) sounding, 0600 G.M.T., 4 September 1958. In this and subsequent tephigrams dry adiabatics are labelled in potential temperature in $^{\circ}\text{C}$, and wet adiabatics in wet-bulb potential temperature in $^{\circ}\text{C}$ (italic figures). Thin pecked lines are for saturation mixing ratios of 1, 4 and 10 g/kg; isobars are labelled in hundreds of mb. The solid line shows the distribution of temperature and the pecked line the distribution of dew-point.

In this sounding the middle level air is considered to have originated in the Saharan adiabatic layer, the lapse-rate between about 580 and 480 mb (C) having been modified by condensation and small-scale (castellanus) convection. The wet-bulb potential temperature in this layer is about 19°C , while lower it is only 17 to 18°C .

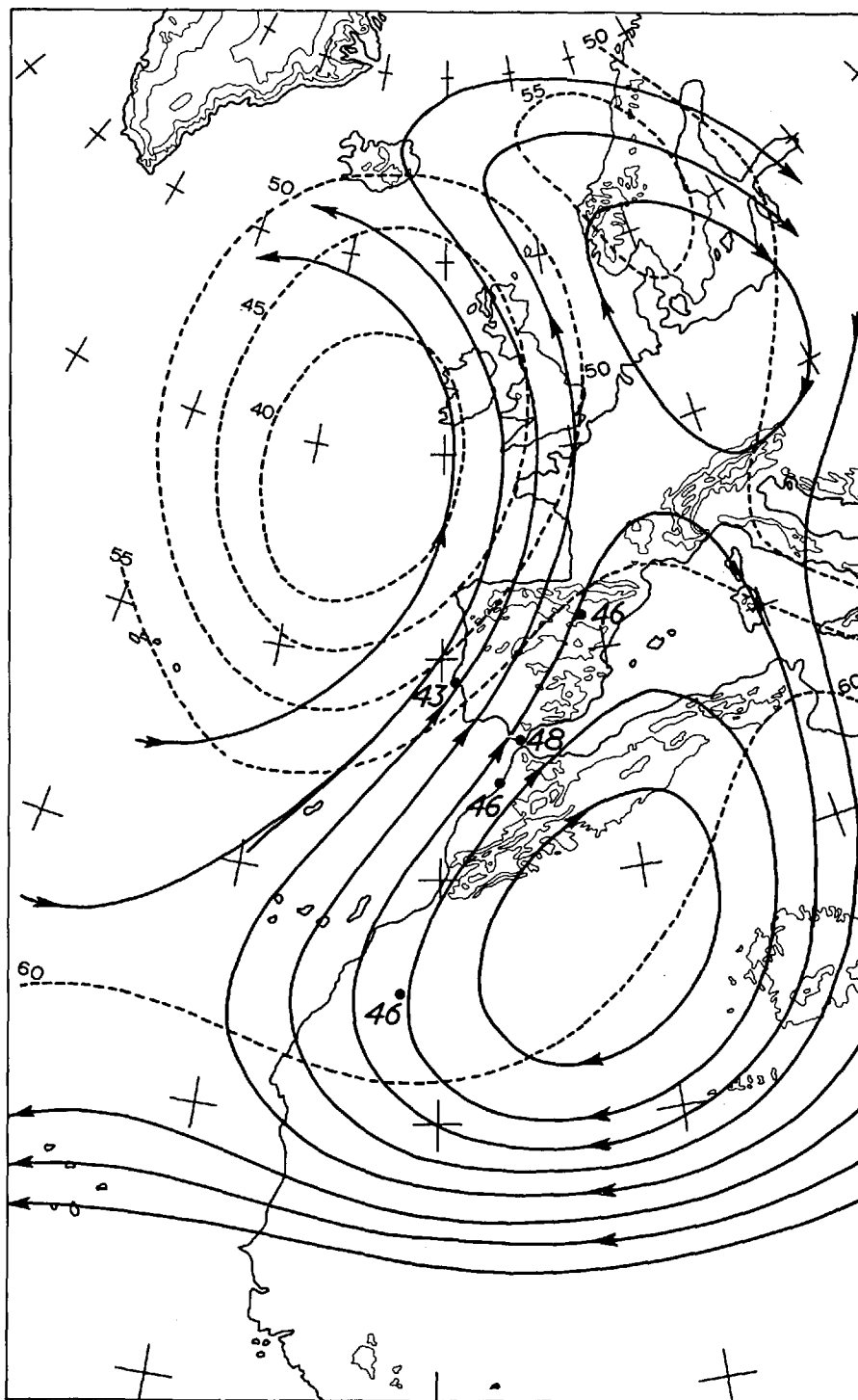


Fig.15. Isentropic relative-flow chart for $\Theta = 43^{\circ}\text{C}$, 1200 G.M.T., 5 September 1968, showing the flow of air from the top of the adiabatic layer over the Sahara into the confluence zone over Spain and Britain. Adiabatic lifting condensation levels are shown in tens of mb (italic figures) beside Fort Trinquet, Kenitra, Gibraltar, Lisbon and Zaragoza.

Fig.16

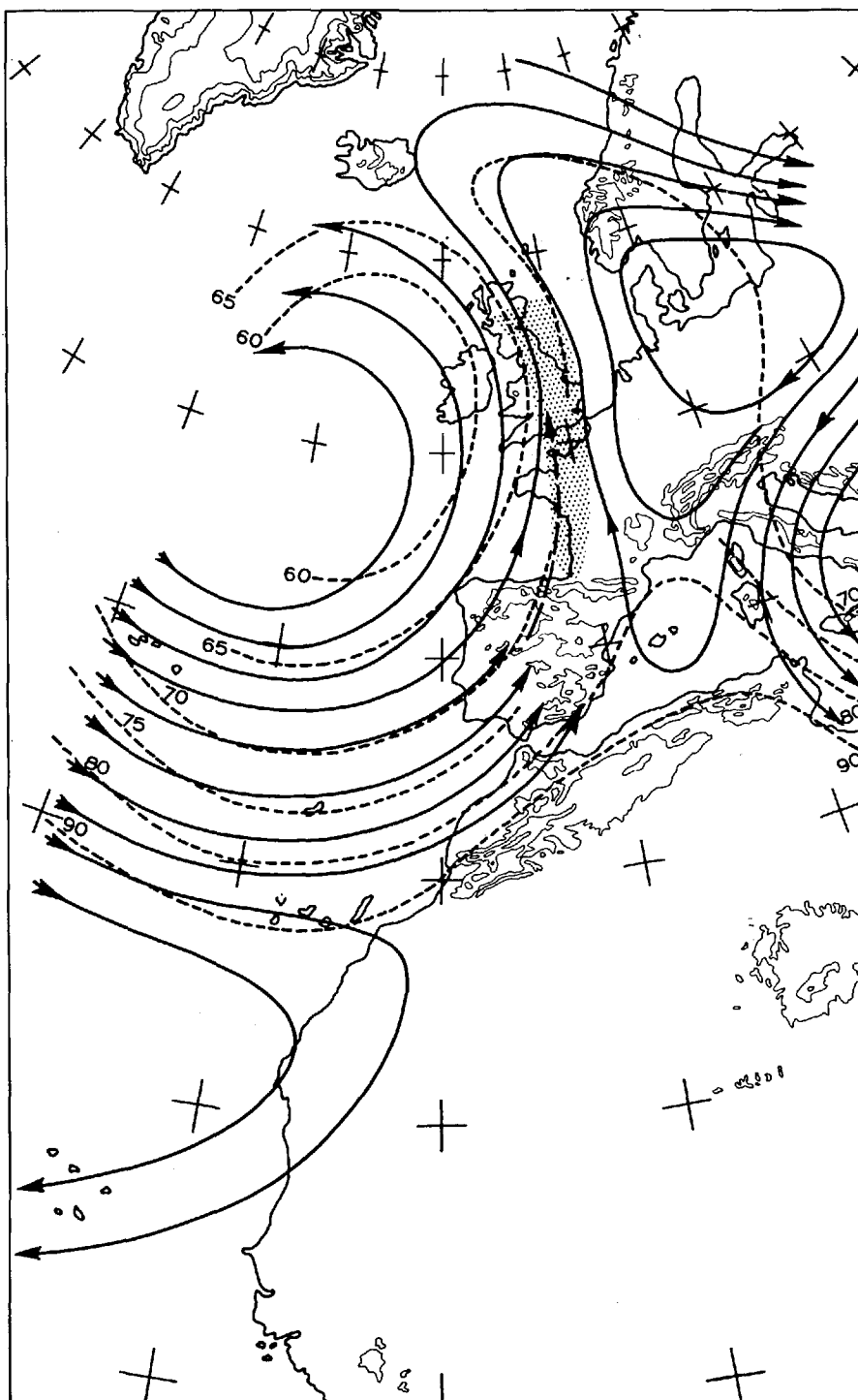


Fig.16. Isentropic relative-flow chart for $\Theta = 33^{\circ}\text{C}$, 1200 G.M.T., 5 September 1958. During the flow over Spain and into the confluence zone farther north, ascent and condensation remove air to higher isentropic surfaces, especially in the belt of extensive altocumulus (with castellanus; stippled area, terminated arbitrarily over Scotland).

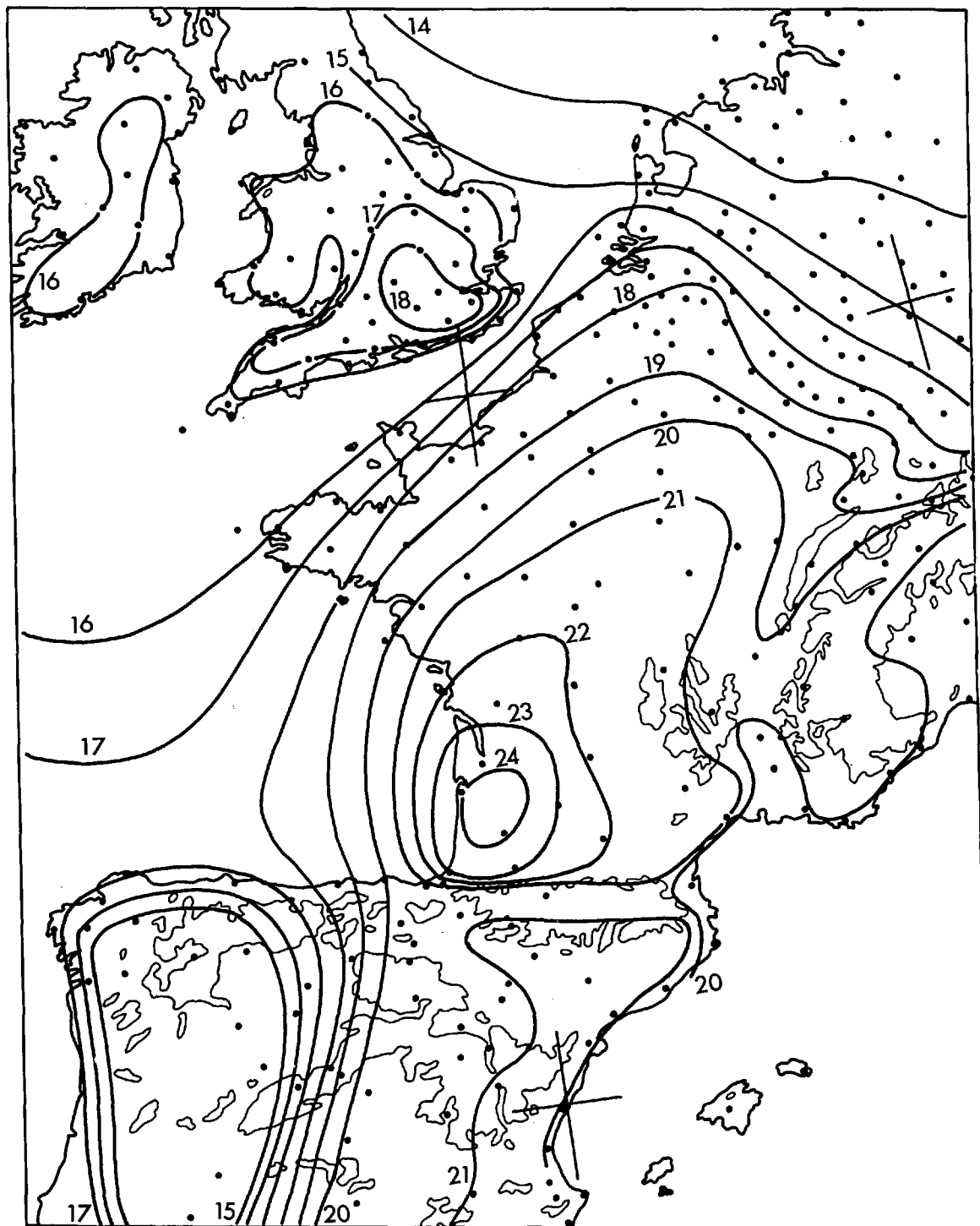


Fig.17. Distribution of screen-level wet-bulb potential temperature (in $^{\circ}\text{C}$) over western Europe, 1500 G.M.T., 4 September 1958. Ground height contours for 3000 ft are drawn as thin lines. Dots show the positions of observing stations. At this time the temperatures were about the maxima 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.

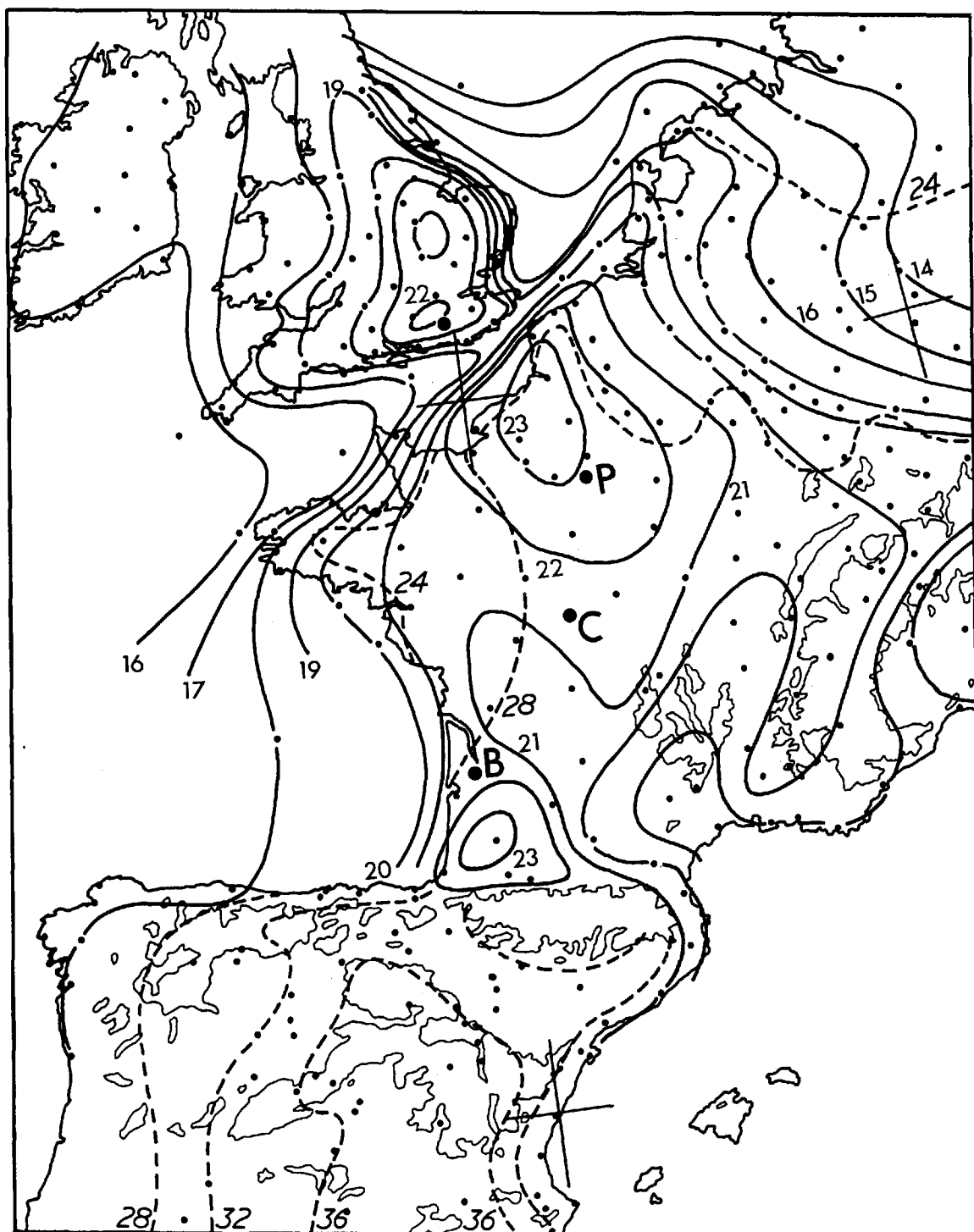


Fig.18. Distribution of screen-level wet-bulb potential temperature (in $^{\circ}\text{C}$), and of ordinary potential temperature (pecked lines labelled in italic figures), over western Europe, 1500 G.M.T., 5 September 1958 (see also Fig.17). The letters P, C and B show the locations of the sounding stations at Paris (Trappes, 07145), Chateauroux (07354) and Bordeaux (07510), and the large dot on the isopleth for 22°C over S.E. England the location of Crawley (03774).

Note that the original area of maximum wet-bulb potential temperature (Fig.17) has during the intervening 24 hours been displaced to Normandy, while another has developed over southern France.

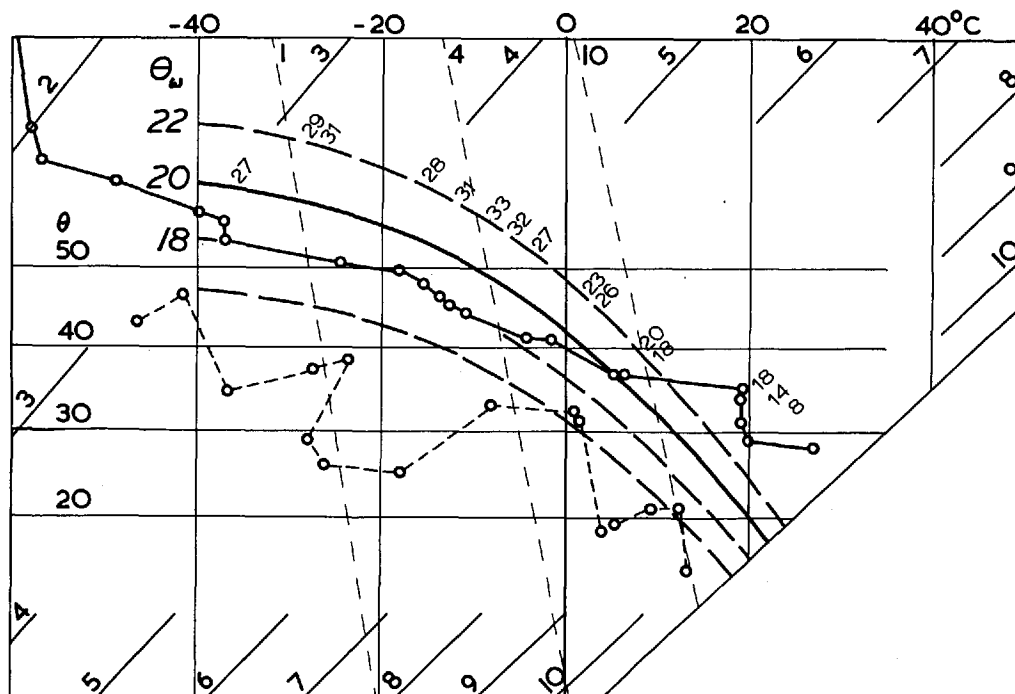


Fig.19. Sounding for Chateauroux (07354), 1200 G.M.T., 5 September 1958, showing air from the Spanish adiabatic layer present between 850 and 700 mb. The lid can be considered to be at 850 mb. Some castellanus were present in the Spanish plume above 700 mb. The small figures to the right of the sounding show the profile of wind speed in kts.

The convection over France is confined below the base of the Spanish plume at about 850 mb, leading to the large rise in wet-bulb potential temperature.

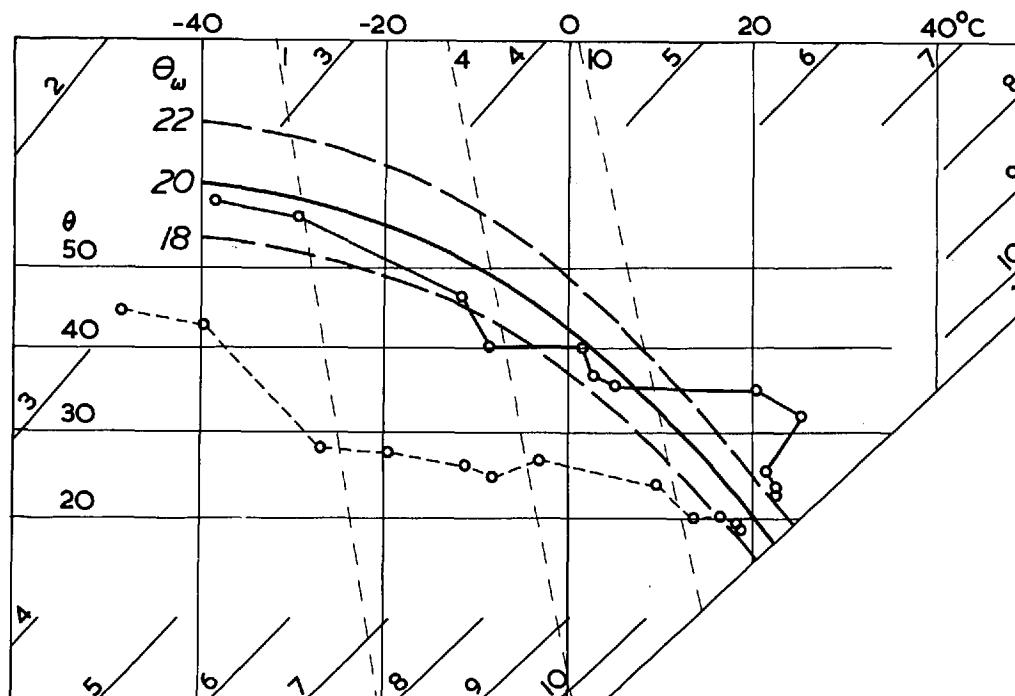


Fig.20. Sounding for Bordeaux (07510), at 000 G.M.T., 5 September 1958. The Spanish plume is present with base at 850 mb and top in the middle troposphere, with some castellanus in its upper part.

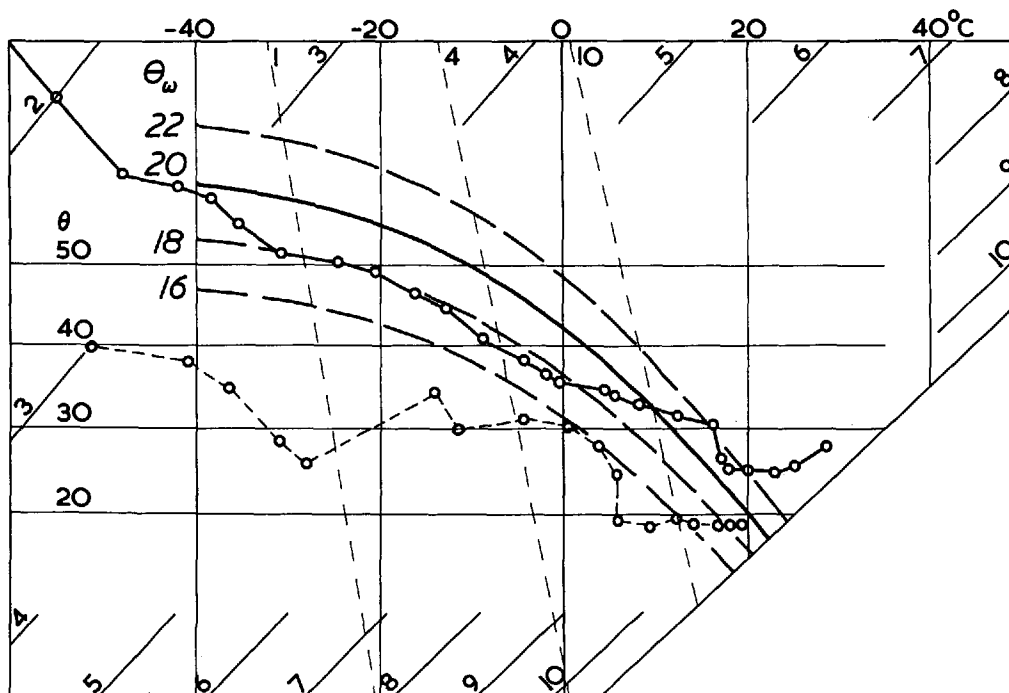


Fig. 21. Sounding for Bordeaux (07510), at 1200 G.M.T., 5 September 1958. The Spanish plume is present above 850 mb, with castellanus and a lapse-rate modified by condensation.

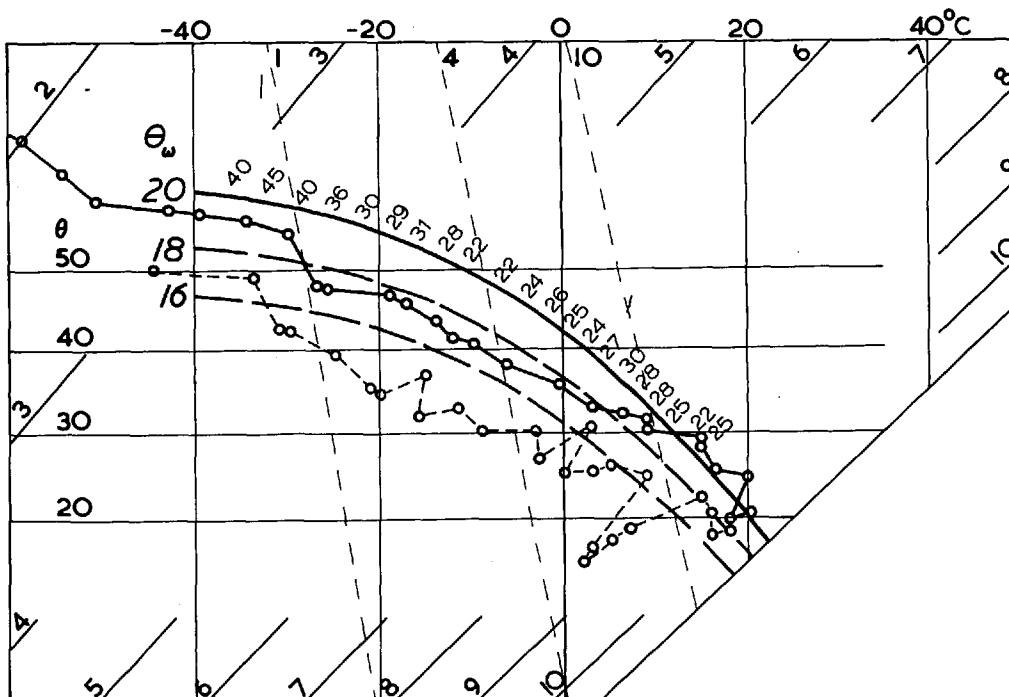


Fig. 22. Sounding at Crawley (03774), 1200 G.M.T., 5 September 1958. The figures to the right of the sounding show the profile of wind speed in kts. The Spanish air is present above 770 mb, but its base has now risen and cooled, so that it cannot restrict the growth of cumulus formed in the French air. The fringe of the abnormally moist French convective layer, previously heated beneath the Spanish plume, is present near 900 mb, where the wet-bulb potential temperature exceeds 19°C .

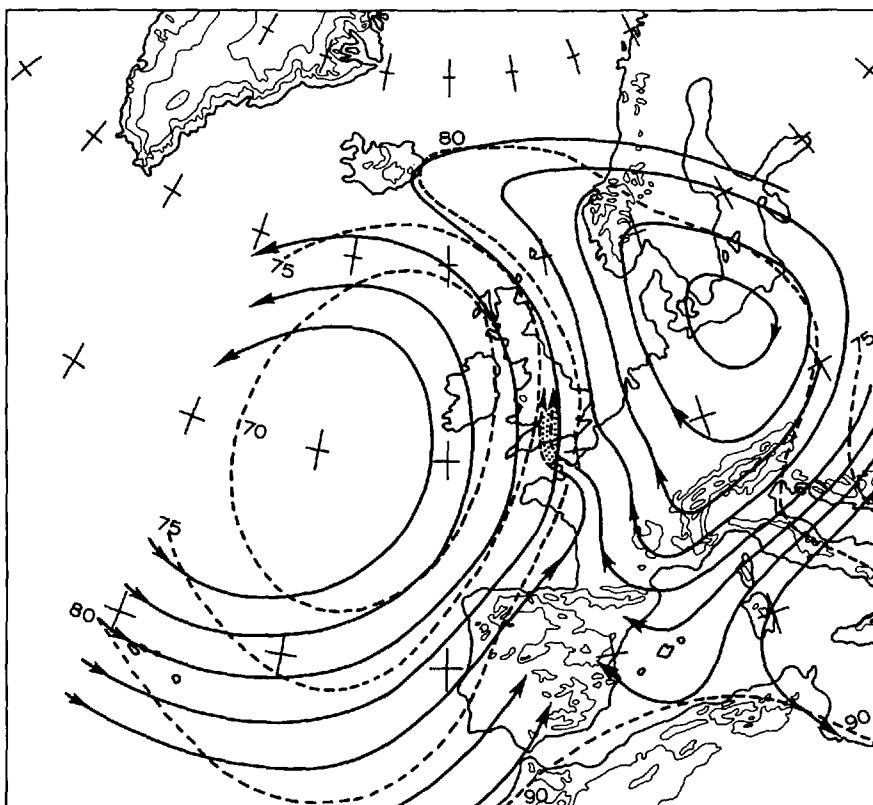


Fig.23. Isentropic relative-flow chart for $\Theta = 28^{\circ}\text{C}$, 1200 G.M.T., 5 September 1958. The chart shows the flow of air from the adiabatic layer over France into the confluence zone. Over the English channel large-scale ascent leads to castellanus formation near or west of the edge of the Spanish plume, and the air leaves the isentropic surface for higher levels (streamlines become pecked, with stipple between).

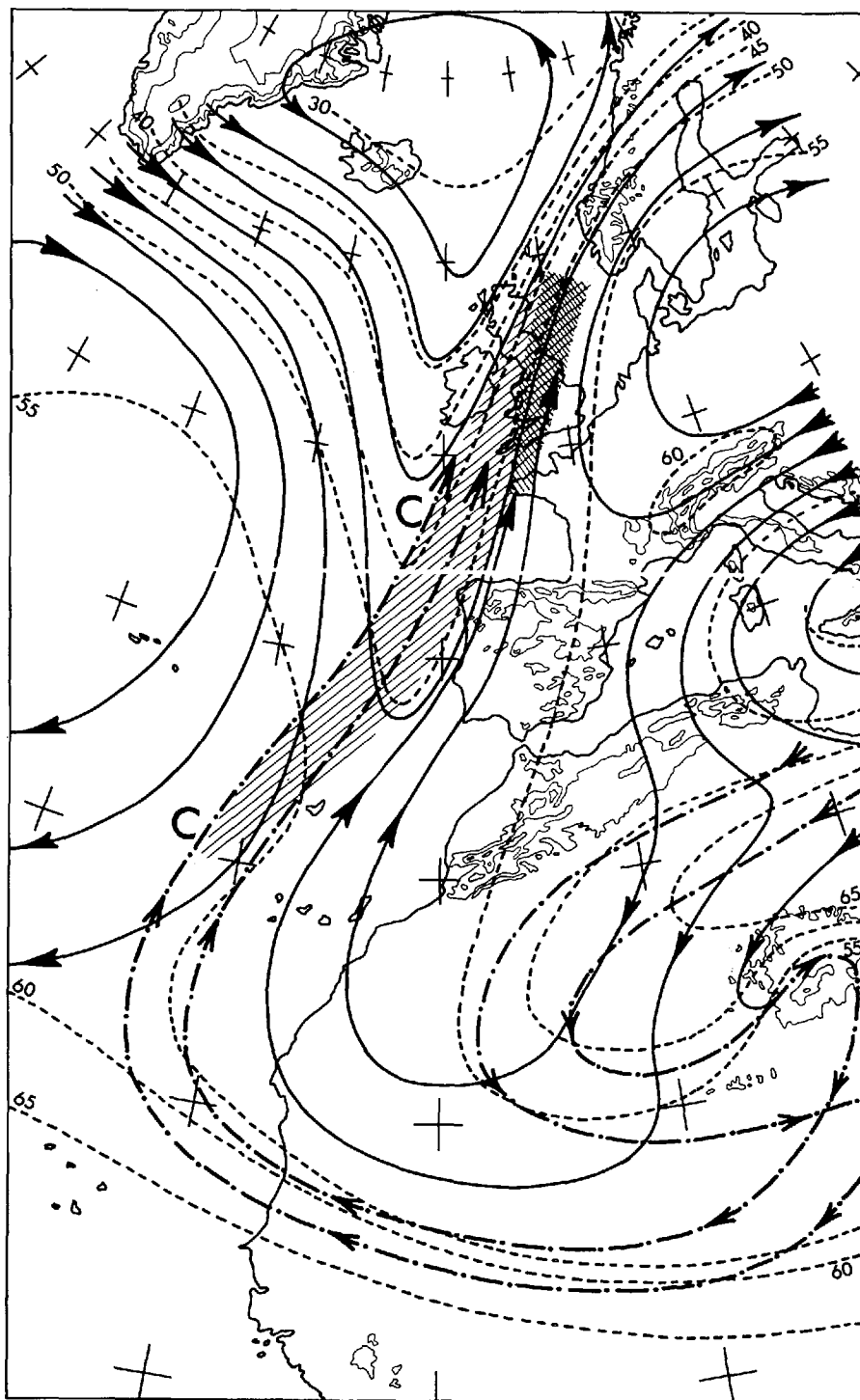


Fig.24. Isentropic relative flow chart for $\Theta = 430^{\circ}\text{C}$, 1200 G.M.T. 9 July 1959. The chart shows the flow of air from the Sahara into the confluence zone west of Spain and over Britain. The streamlines over the Sahara show the flow at the top of the adiabatic layer; additionally, the dot-dash lines show the surface flow, and the flow above the adiabatic layer away from the Ahaggar mountains which forms the southern fringe of the flow out of the Sahara, and which rises above this isentropic chart where condensation in it produces a cirrus belt (open hatching) whose western edge lies along the streamline marked CC. The adiabatic layer from the Sahara produces an alto-cumulus belt (cross-hatching- both cloud belts are arbitrarily terminated).

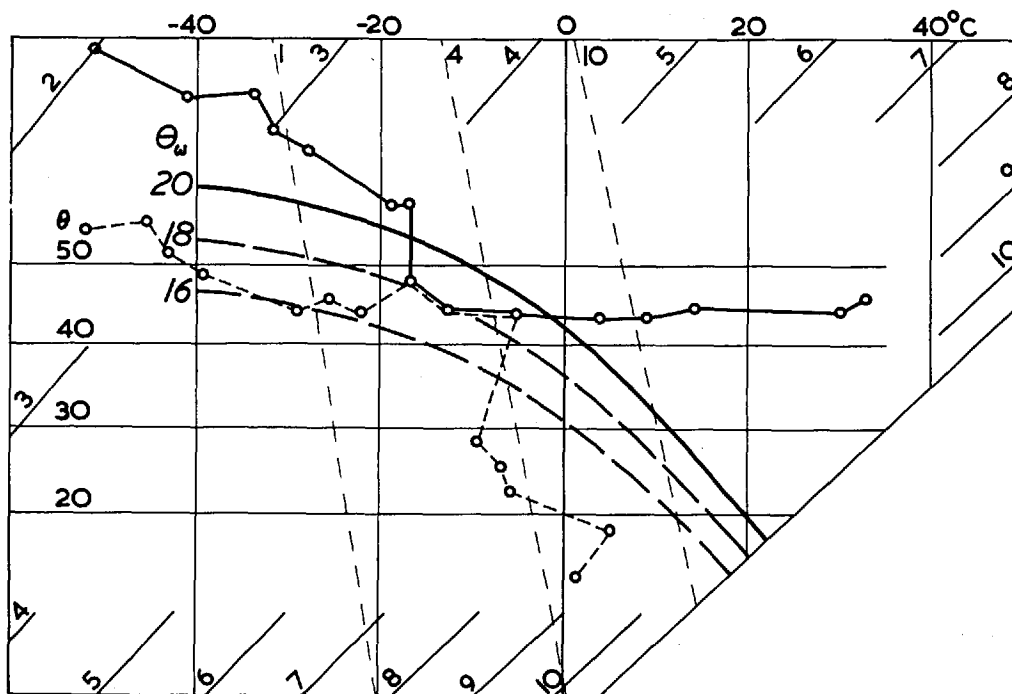


Fig.25. Sounding for Tamanrasset (60880), 1200 G.M.T., 9 July 1959. The air is reported as saturated between 550 and 450 mb, at the top of the adiabatic layer.

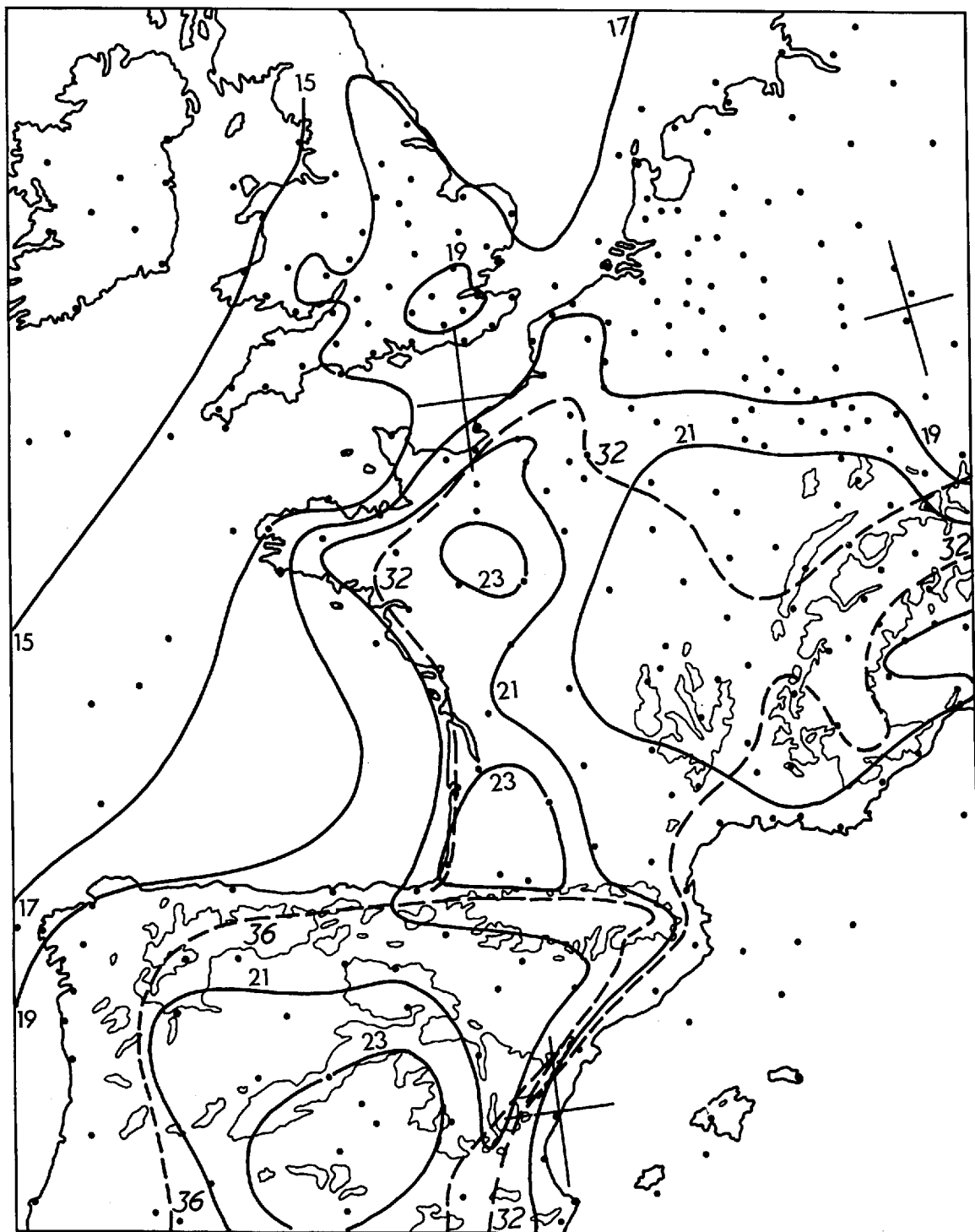


Fig.26. Distribution of screen-level potential temperatures over western Europe at 1500 G.M.T., 8 July 1959 (wet-bulb potential temperatures: solid lines labelled at intervals of 20°C; ordinary potential temperatures: pecked lines at intervals of 4°C). The characteristic region of high wet-bulb potential 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.

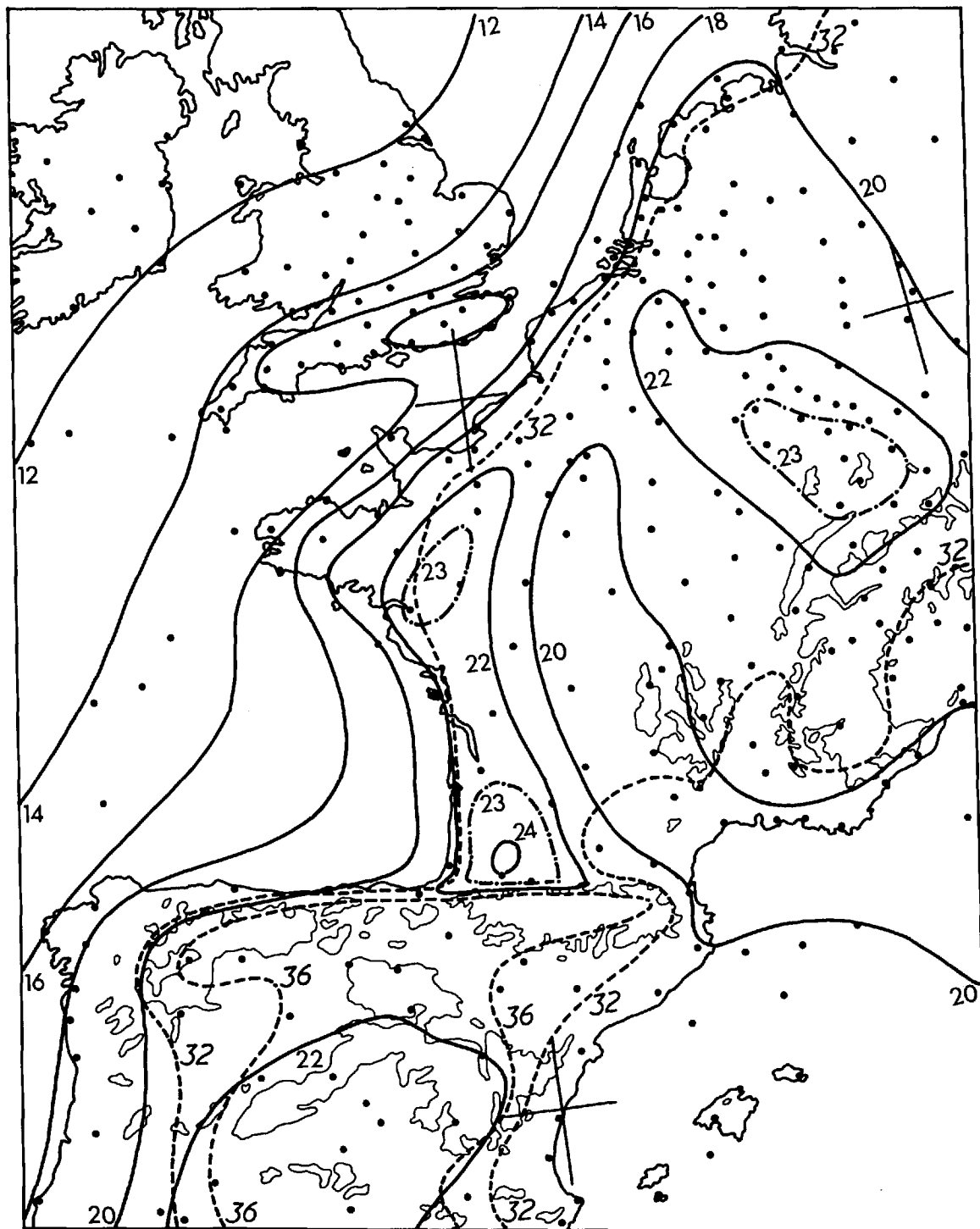


Fig.27. Distribution of screen-level potential temperatures over western Europe at 1500 G.M.T., 9 July 1959 (wet-bulb potential temperatures: solid lines labelled at intervals of 2°C ; ordinary potential temperatures: pecked lines labelled 32 and 36°C). The pattern over France is similar to that on the previous day (Fig.26), and the moist area in the north-west sustained a further outbreak of severe storms over the Channel and southern England during the night.

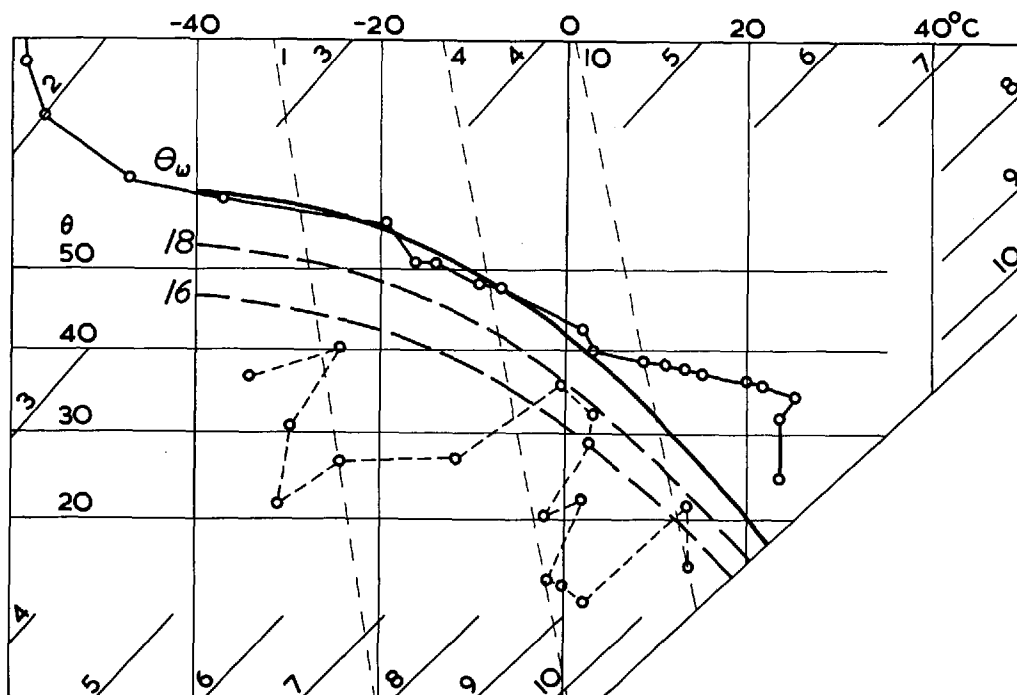


Fig.28. Sounding at Zaragoza (08159), 0000 G.M.T., 9 July 1959. The small-scale convection during the previous day appears to have extended generally up to 650 mb, and perhaps locally over high ground to 450 mb or even higher.

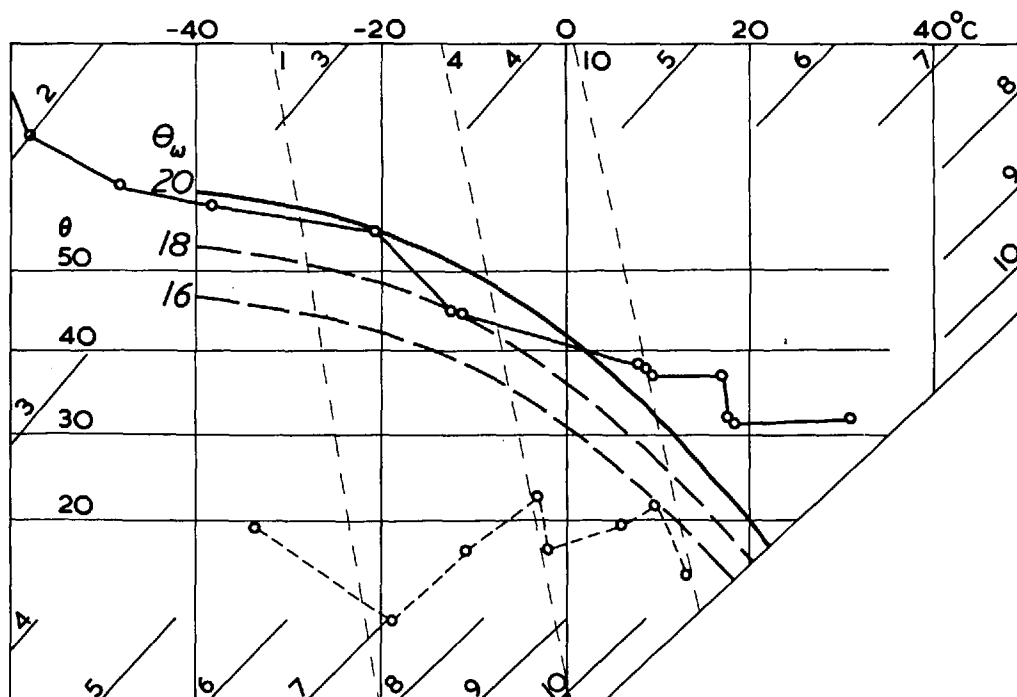


Fig.29. Sounding at Chateauroux (07354), 1200 G.M.T., 9 July 1959. The base of the Spanish plume which constitutes the lid is at about 800 mb, and the top probably at 700 mb: above this level the air becomes very dry.

Fig.30

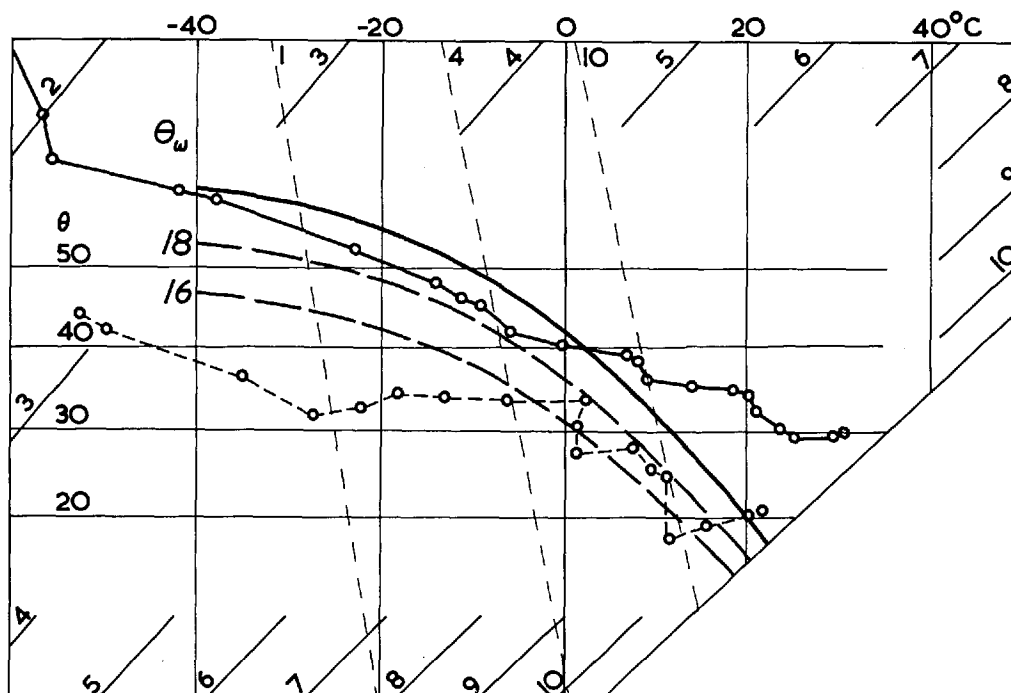


Fig.30. Sounding at Bordeaux (07510), 1200 G.M.T., 9 July 1959. The base of the Spanish plume is at about 850 mb, and condensation has probably occurred around 700 mb. The air between 580 and 680 mb appears to have originated in the adiabatic layer over the Sahara.

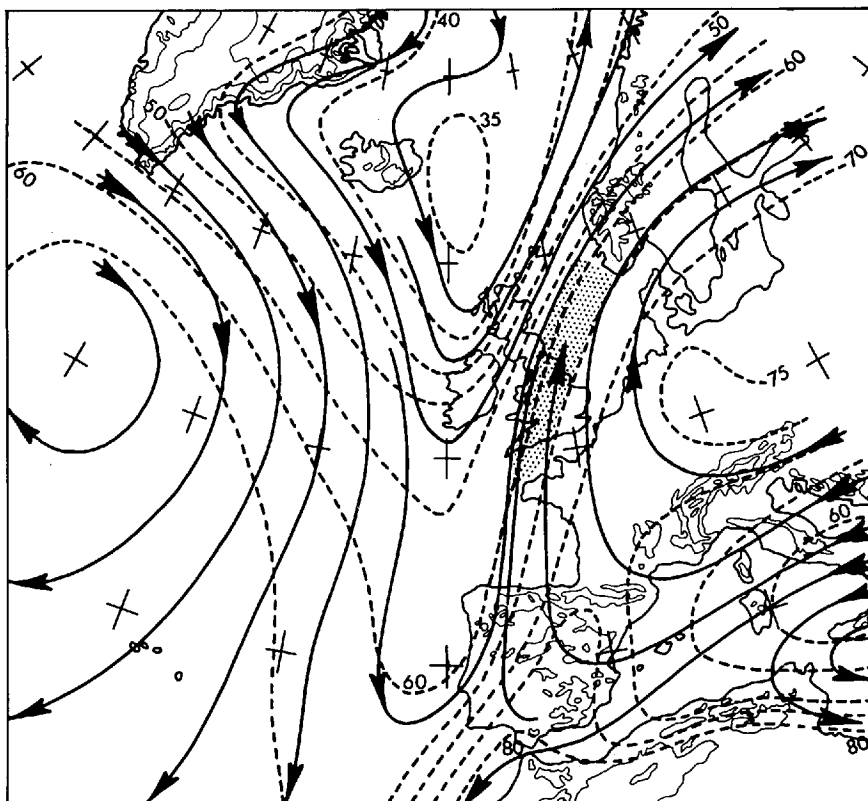


Fig. 31. Isentropic relative flow chart for $\Theta = 37^{\circ}\text{C}$, 1200 G.M.T., 9 July 1959. The pecked isobars are drawn at intervals of 50 mb. The air which flows across Spain enters the confluence zone farther north and ascends to produce castellanus clouds over S.W. England (stippling, arbitrarily terminated over the North Sea).

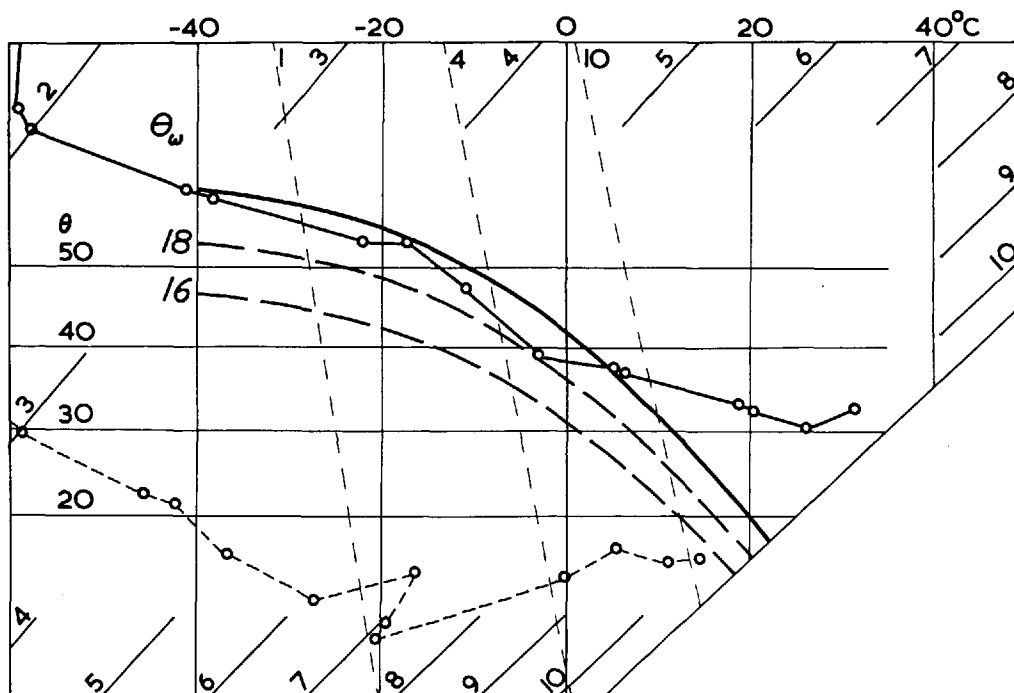


Fig. 32. Sounding for Trappes (07145), 1200 G.M.T., 9 July 1959. There is no Spanish plume, and the air in the lowest layers is very dry.

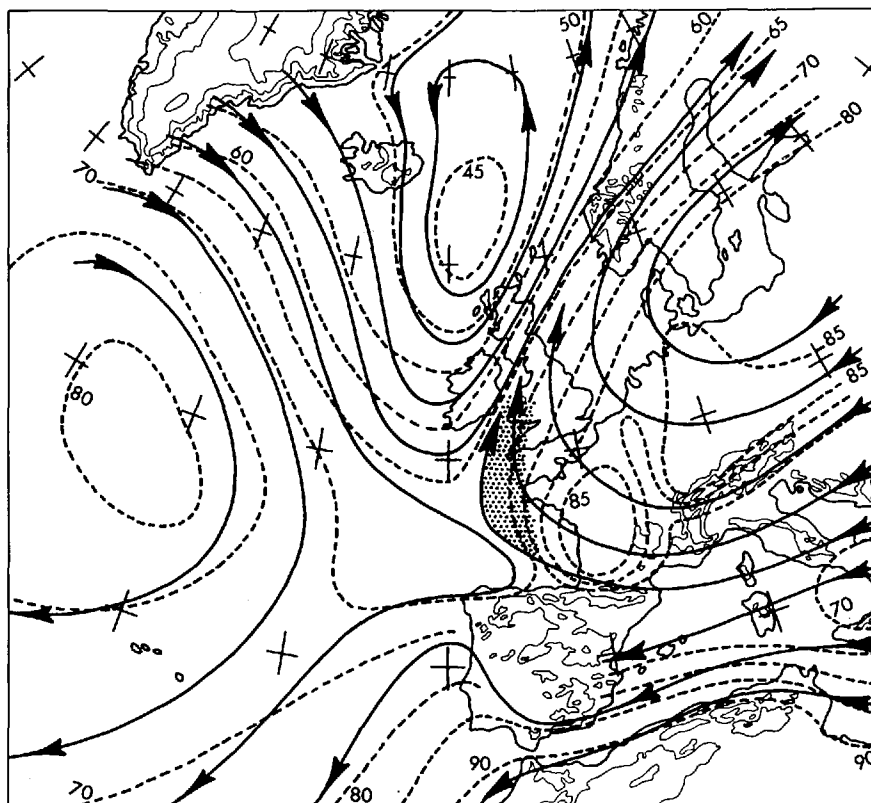


Fig. 33. Isentropic relative flow chart for $\Theta = 33^{\circ}\text{C}$, 1200 G.M.T., 9 July 1959. The chart shows the flow of air from the French adiabatic layer into the confluence zone, with ascent and castellanus formation in the stippled belt (arbitrarily terminated over Wales).

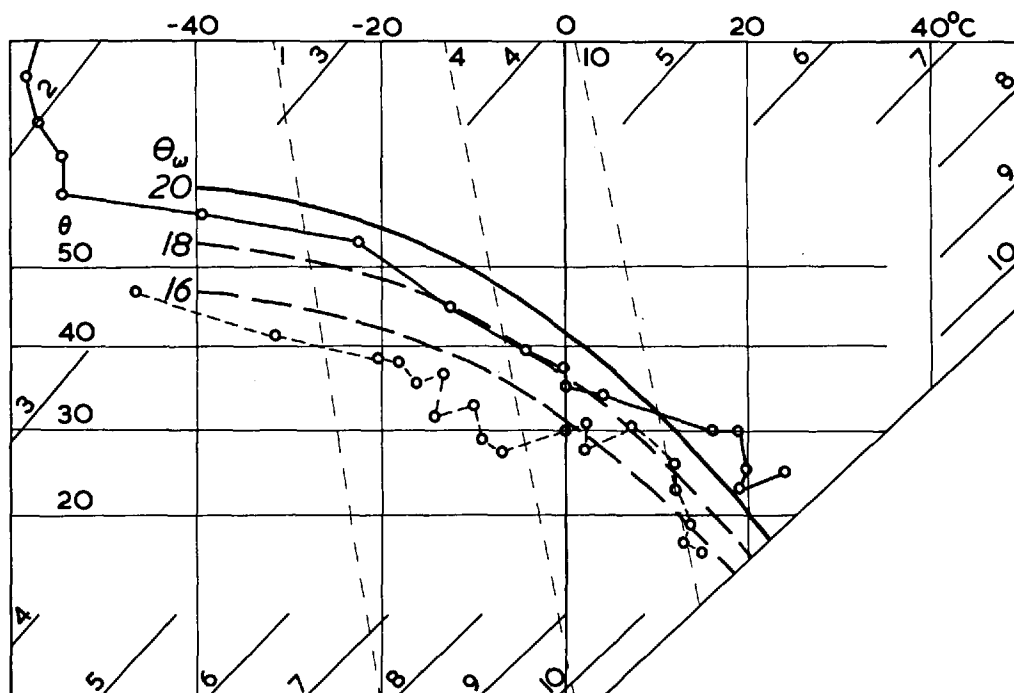


Fig. 34. Sounding for Crawley (03774), 1200 G.M.T., 9 July 1959. The fringe of the abnormally moist French convective layer, previously heated beneath the lid formed by the Spanish plume, is present near 850 mb, where the wet-bulb potential temperature is 20°C.

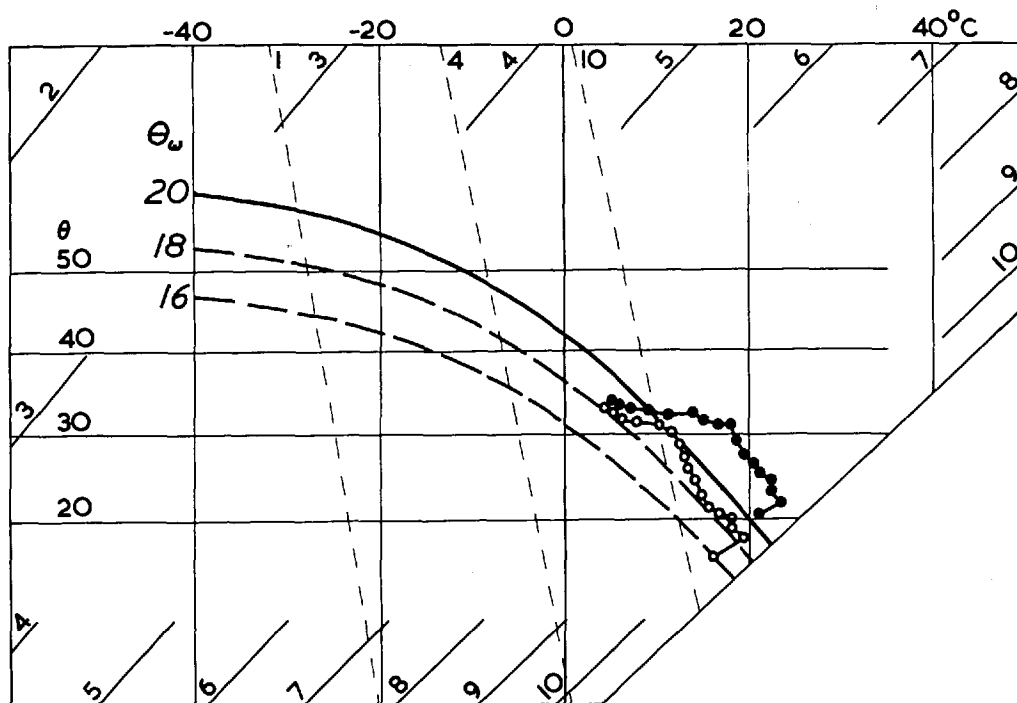


Fig. 35. Sounding for Shoeburyness (03693, 50 miles N.E. of Crawley), 1200 G.M.T., 9 July 1959. The full circles show the distribution of temperature, the open ones that of wet-bulb temperature. As on the Crawley sounding (Fig. 34), the abnormally moist French convective layer is present above 850 mb.

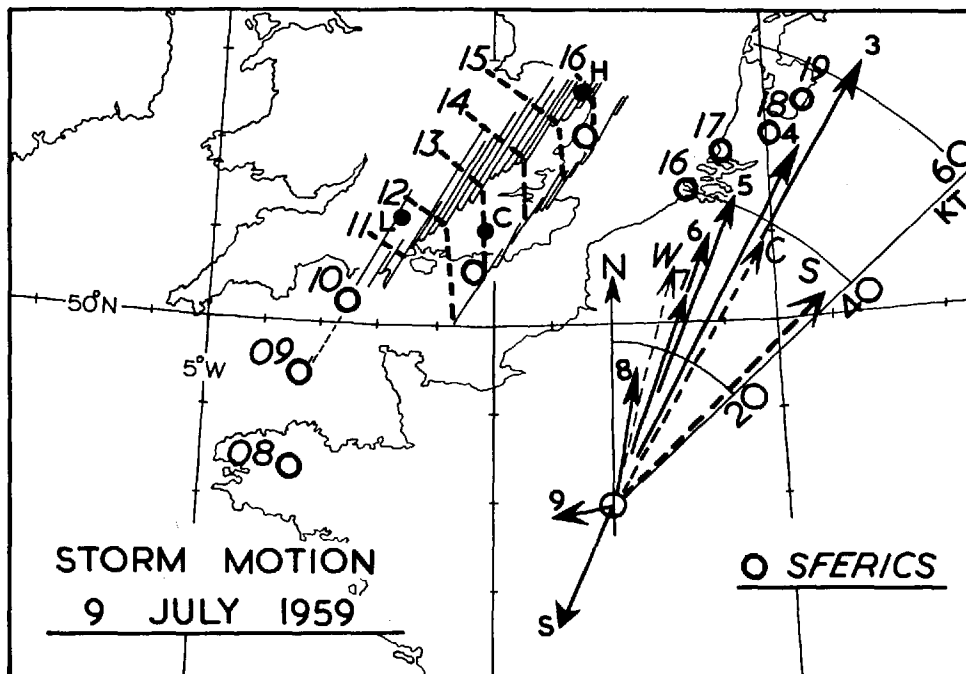


Fig.36. Origin and movement of the Wokingham storm, 9 July 1959. The open circles show the hourly positions of intense sferic clusters, beginning over Brittany at 08 hrs G.M.T. The thin lines over England are the paths of individual radar echo cells of the storm, successively replaced on its right flank, and the thick pecked lines show their position at each hour. The letter C marks the sounding station at Crawley.

On the right of the diagram are drawn vectors representing the winds near the storm at the surface (S) and at each hundred millibar level to 300 mb, the velocity of weak echo cells (W), of the intense cells (*italic C*) and of the storm (*italic S*).

Note that on the squall front of the Wokingham storm a second severe storm develops over the extreme south-east of England, and later another near the Dutch coast.

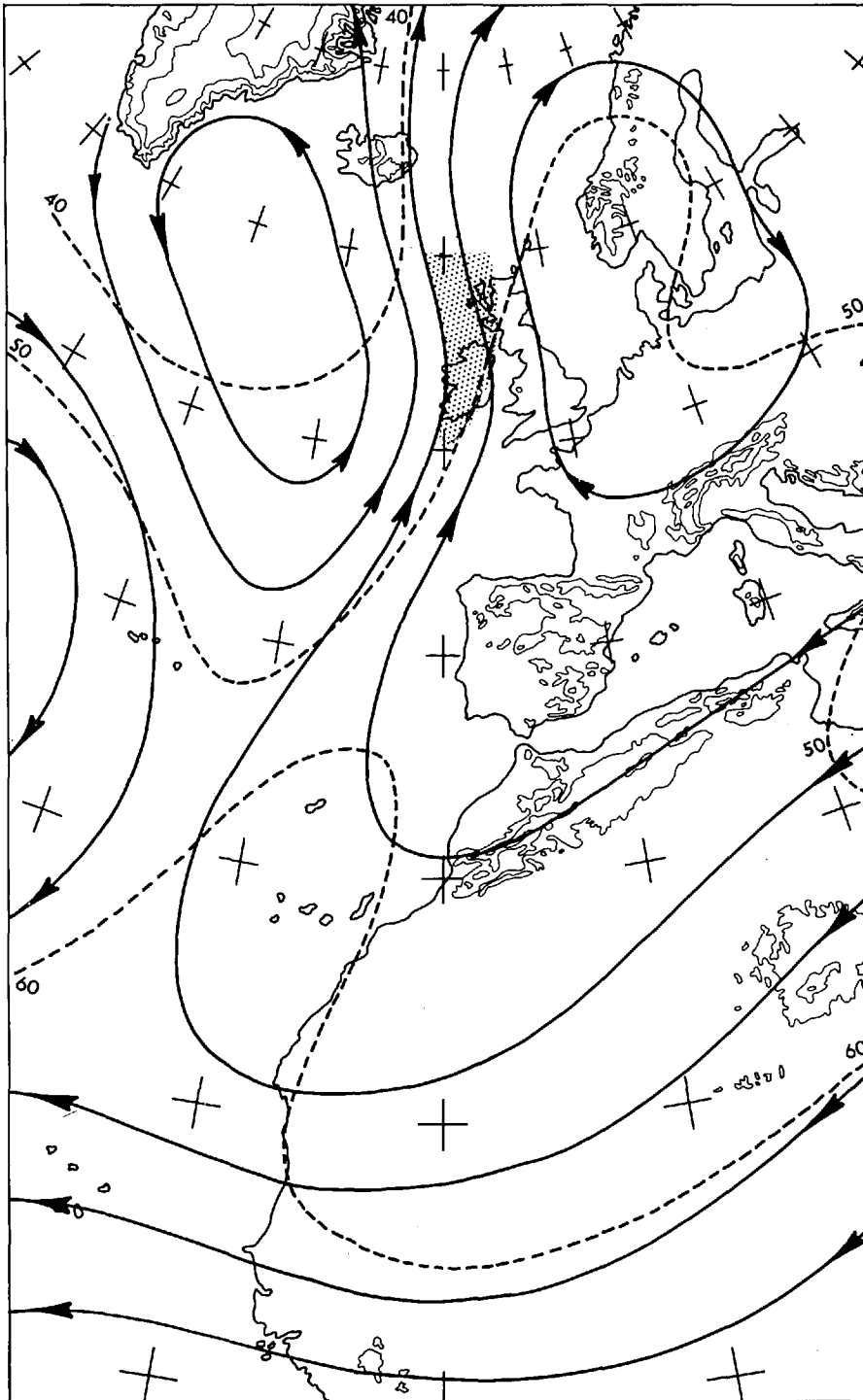


Fig. 37. Isentropic relative flow chart for potential temperature 430°C, 1200 G.M.T., 29 August 1961. The flow from the Sahara into the confluence zone west of Britain produced a belt of high altocumulus and cirrus over and north of Ireland (stipple).

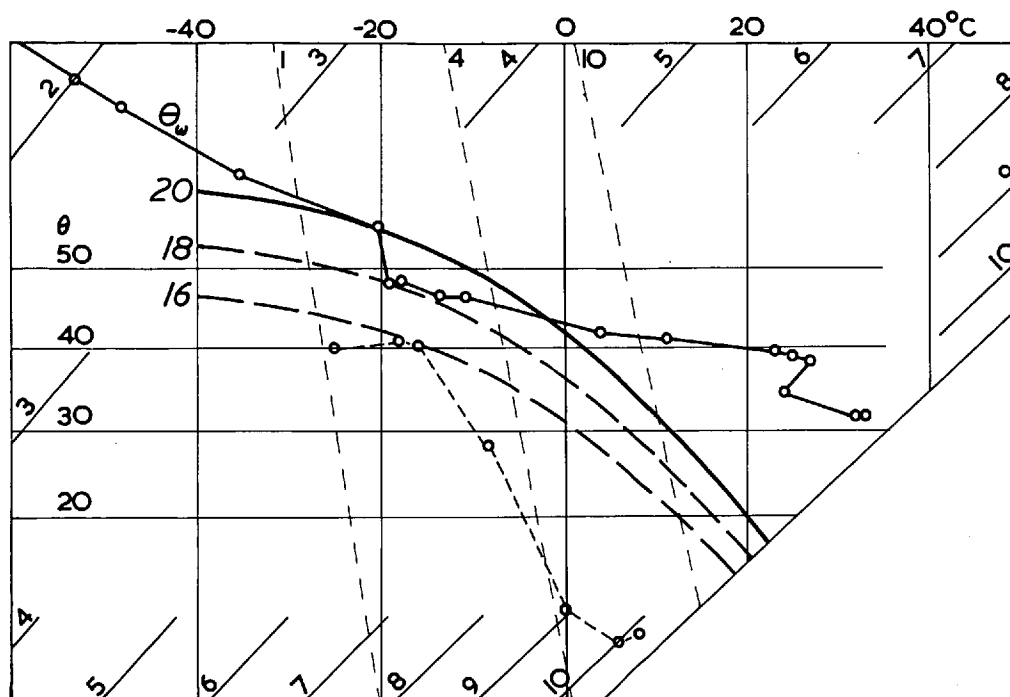


Fig. 38. Sounding for Santa Cruz (60020), 0000 G.M.T., 29 August 1961. The base of the Saharan air is between 880 and 850 mb, and a top between 500 and 450 mb. In the upper part of the layer the lapse-rate has been modified by condensation.

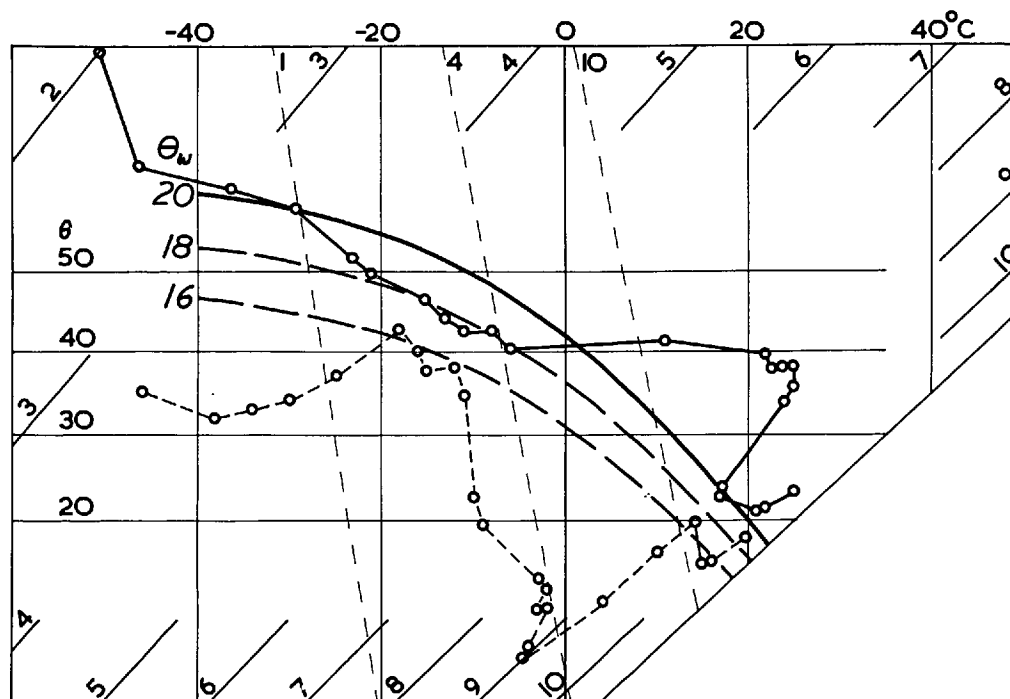


Fig. 39. Sounding for Gibraltar (08495), 1200 G.M.T., 29 August 1961. Saharan air is present between 820 and about 470 mb, with castellanus at the top of the layer (cf. Fig. 38).

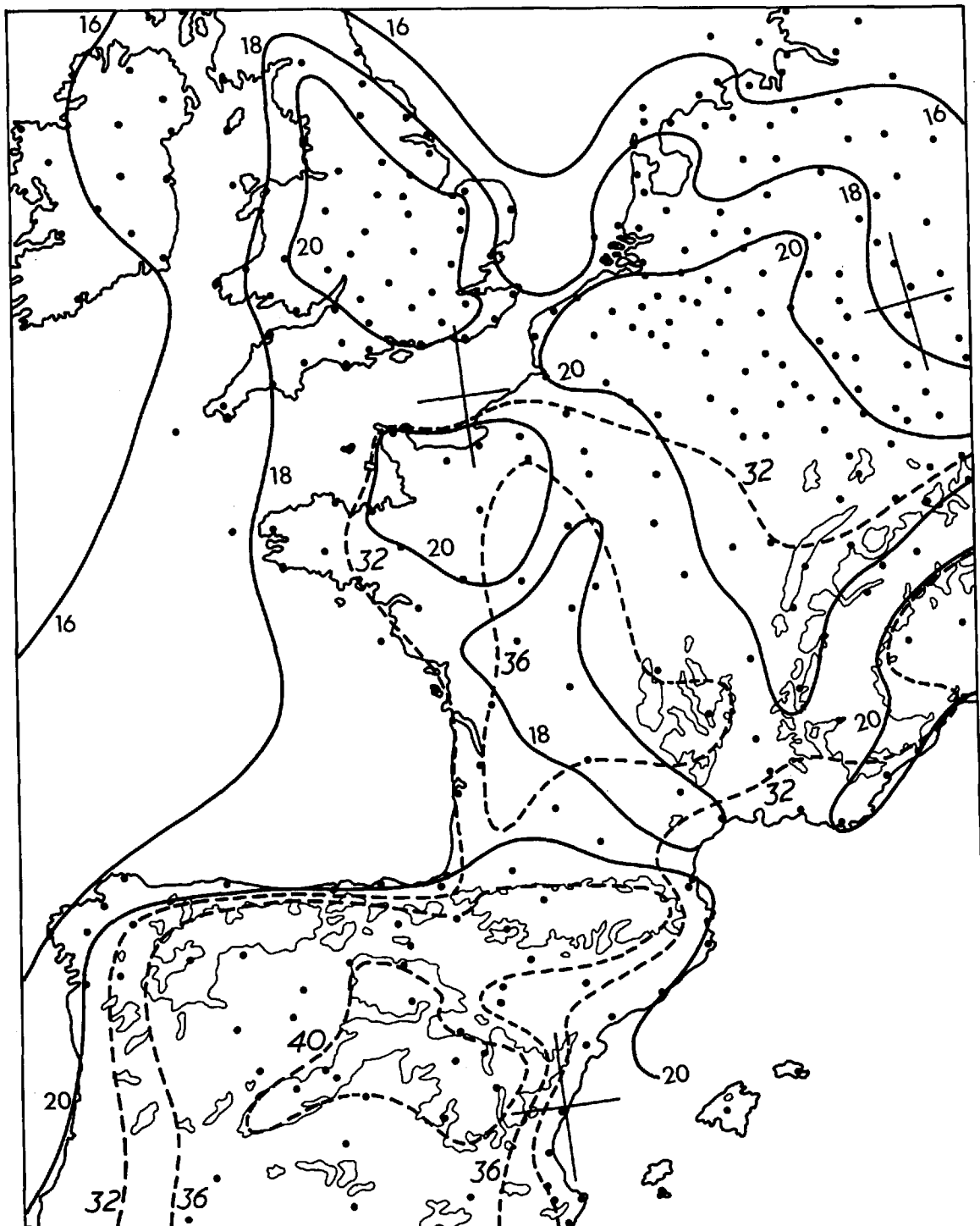


Fig. 40. Distribution of wet-bulb potential temperature (solid lines, labelled in even degrees Centigrade) and ordinary potential temperature (pecked lines at intervals of 4°C), 1500 G.M.T., 29 August 1961. The diagram shows (over France) abnormally high values of the latter, and low values of the former, compared, for example, with the distributions shown in Figs. 18 and 27.

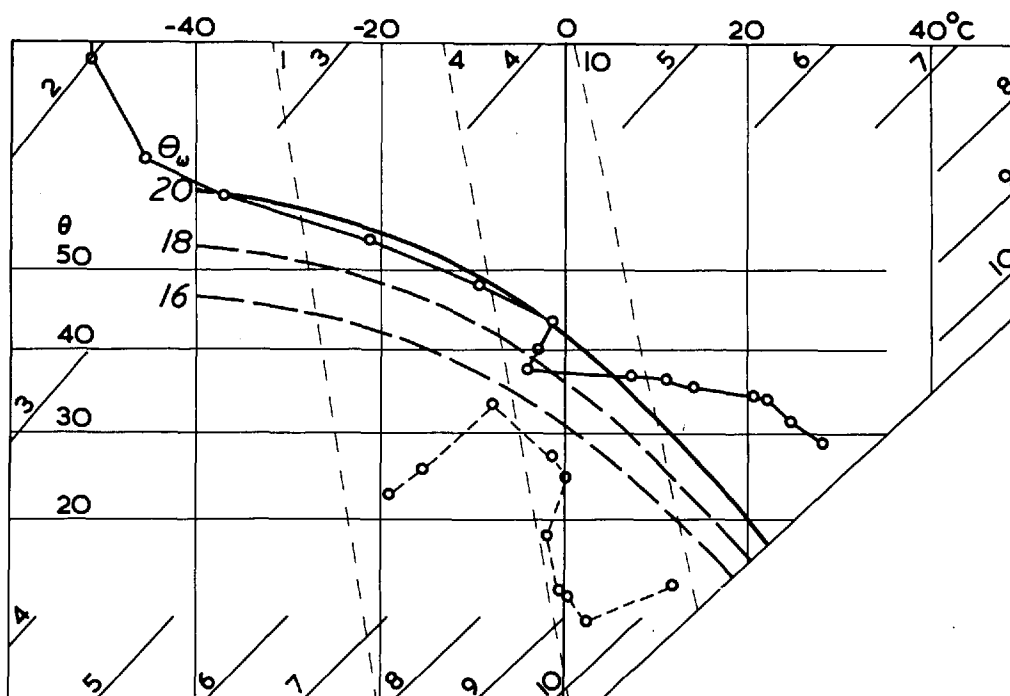


Fig.41. Sounding for Zaragoza, 1200 G.M.T., 29 August 1961. The station is in a valley, and the local potential temperature has not yet risen to the value of 36 to 37°C in the adiabatic layer arriving aloft from higher ground.

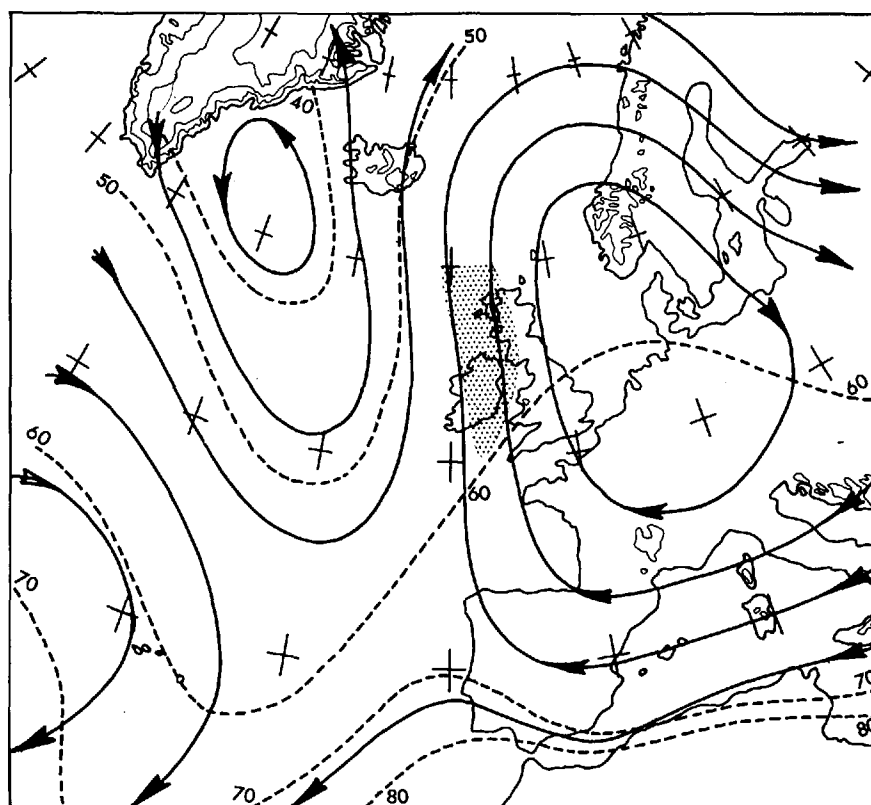


Fig.42. Isentropic relative flow chart for potential temperature 380°C, 1200 G.M.T., 29 August 1961. The pecked lines are isobars at 100 mb intervals. The chart shows the flow of Spanish air into the confluence zone farther north, with the formation of a belt of altocumulus cloud north of the 50th parallel.

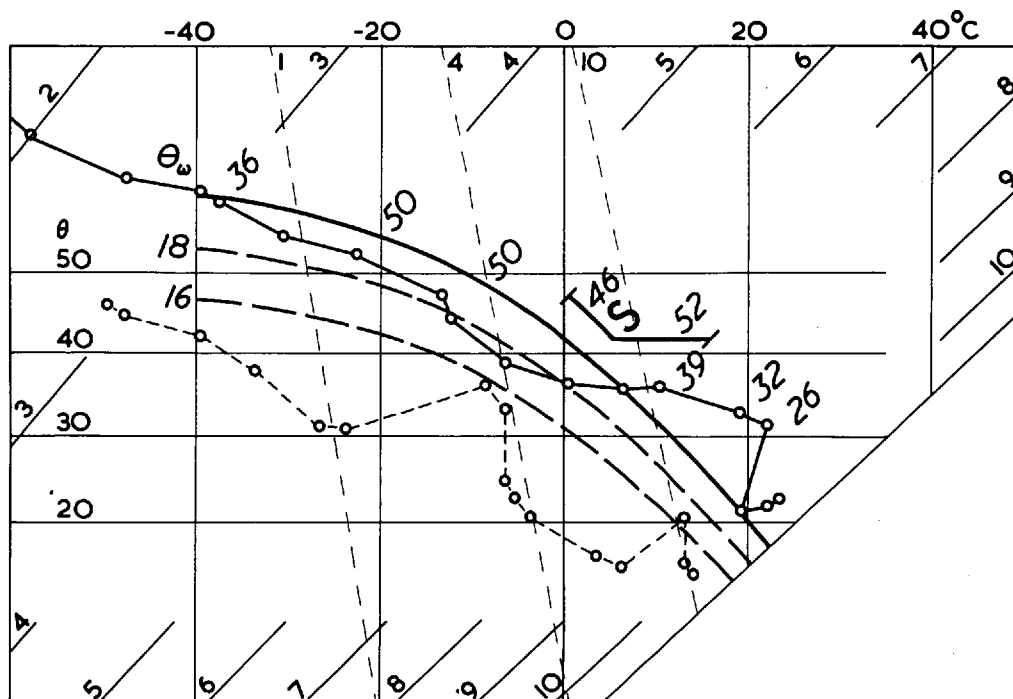
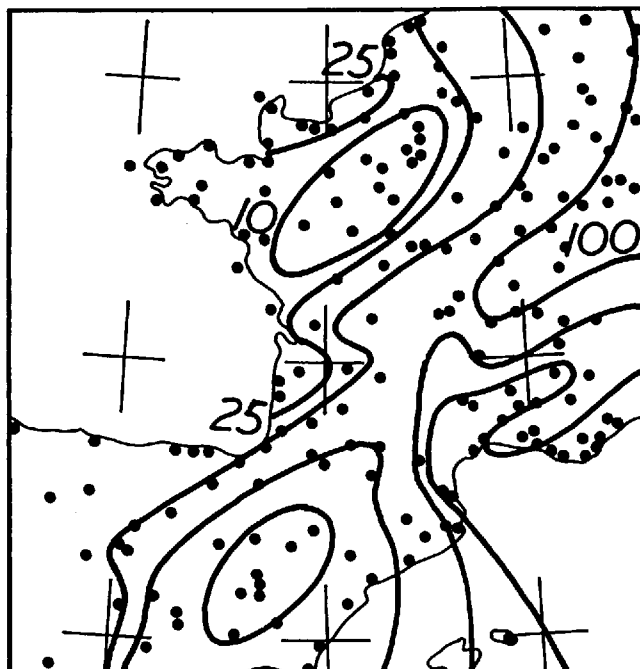


Fig.43. Sounding for Camborne, 1200 G.M.T., 29 August 1961. Spanish air (S) is present with base at 740 mb and top at about 570 mb (cf. Fig.41); the lapse-rate in the upper part of the layer has been modified by condensation. The air immediately below 740 mb probably comes from the French adiabatic layer, but on this occasion is barely distinguishable. The italic figures to the right of the sounding show the profile of wind speed in kts.



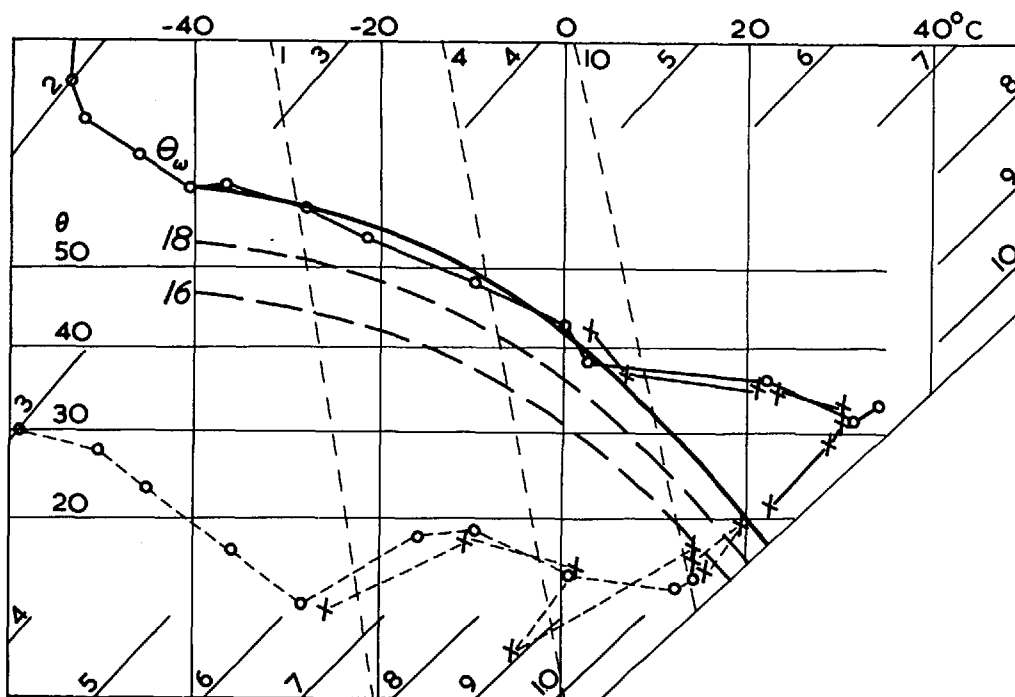


Fig. 45. Soundings for Bordeaux (07510) at 00 hrs (significant points marked by crosses) and at 12 hrs G.M.T. (significant points marked by circles), 29 August 1961. On both soundings the layer between about 850 and 700 mb with a nearly adiabatic lapse-rate marks the plume from the Spanish table-land.

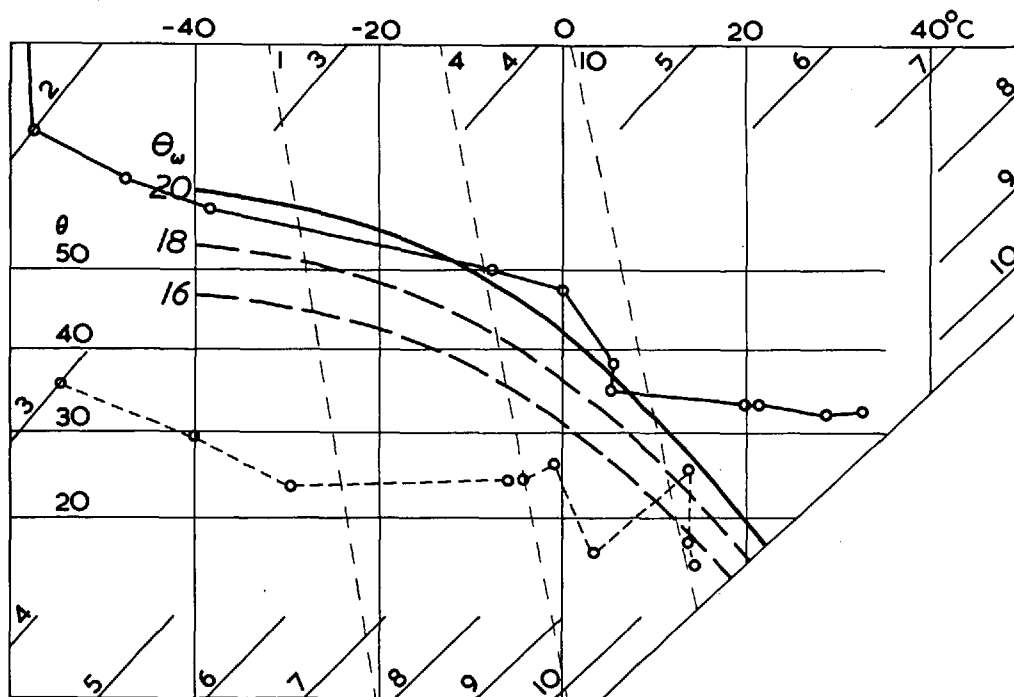


Fig. 46. Sounding for Chateauroux (07354), 1200 G.M.T., 29 August 1961. At this time the local convection is thought to have extended up to 860 mb, and a little later to have penetrated up through the layer of Spanish air to about 700 mb.

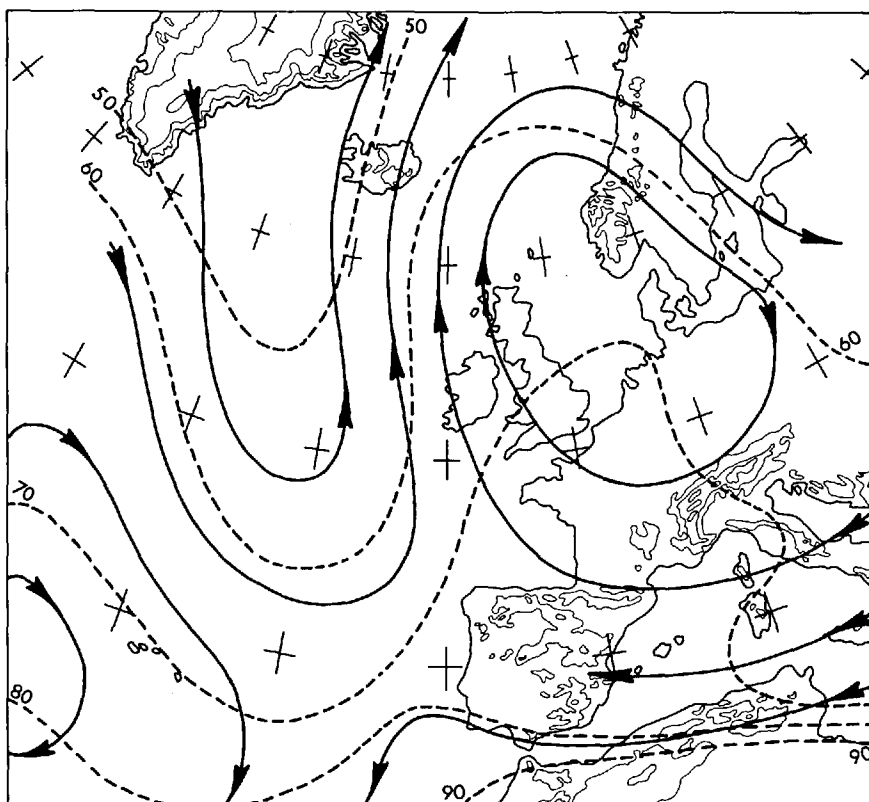


Fig.47. Isentropic relative flow chart for potential temperature 35°C, 1200 G.M.T., 29 August 1961. The pecked lines are isobars at 100 mb intervals. The chart shows the flow from the French adiabatic layer into the confluence zone west of Britain without the formation of cloud.

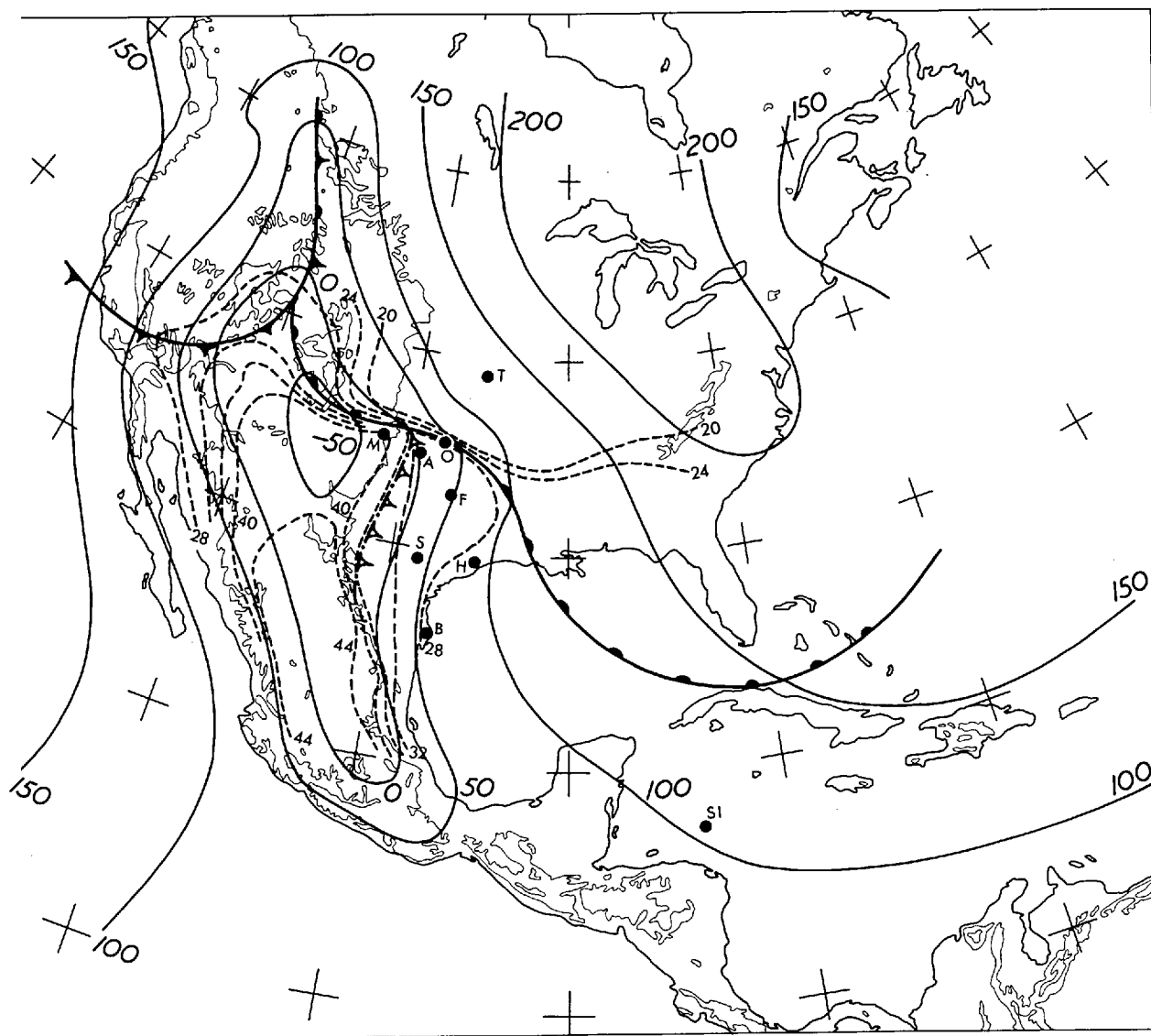


Fig.48. Distribution of screen-level potential temperature (pecked lines labelled at intervals of 4°C) at 2330 G.M.T. on 4 May 1961 (corresponding approximately to the day's maximum values), and conventional frontal analysis with dry-line indicated as a dot-dash line with open cold-front symbols. The continuous lines are 1000 mb contours labelled in m. Letters beside full circles mark the positions of the following places 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, Brownsville; SI, Swan Island.

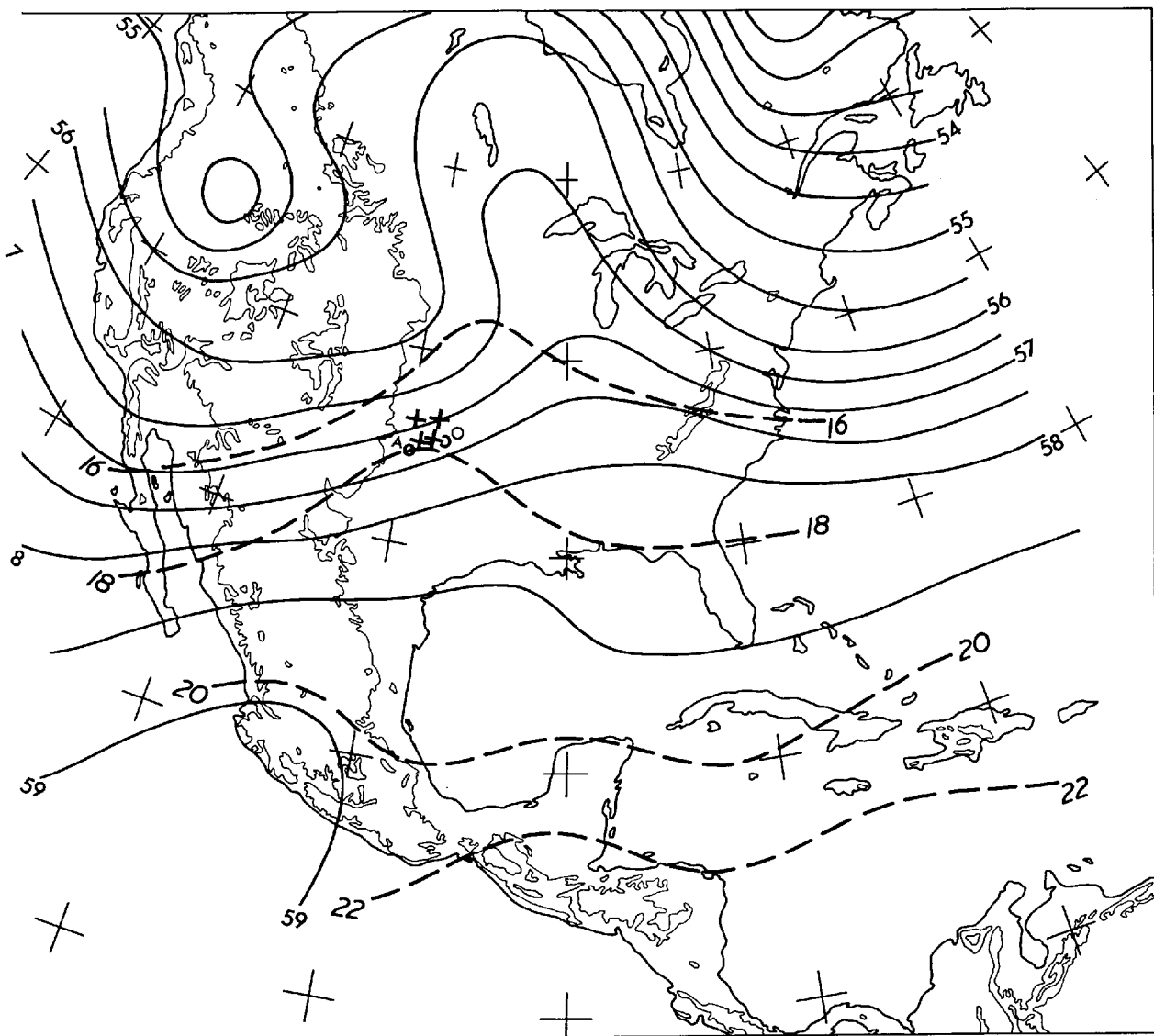


Fig.49. 500 mb contours at 50 m intervals, labelled in hundreds of m, 2330 G.M.T., 4 May 1961. The pecked lines are isopleths of Θ_s , labelled in $^{\circ}\text{C}$. The crosses between Altus (A) and Oklahoma City (O) mark the places from which giant hail or tornadoes were reported between 1500 and 1700 hrs local time.

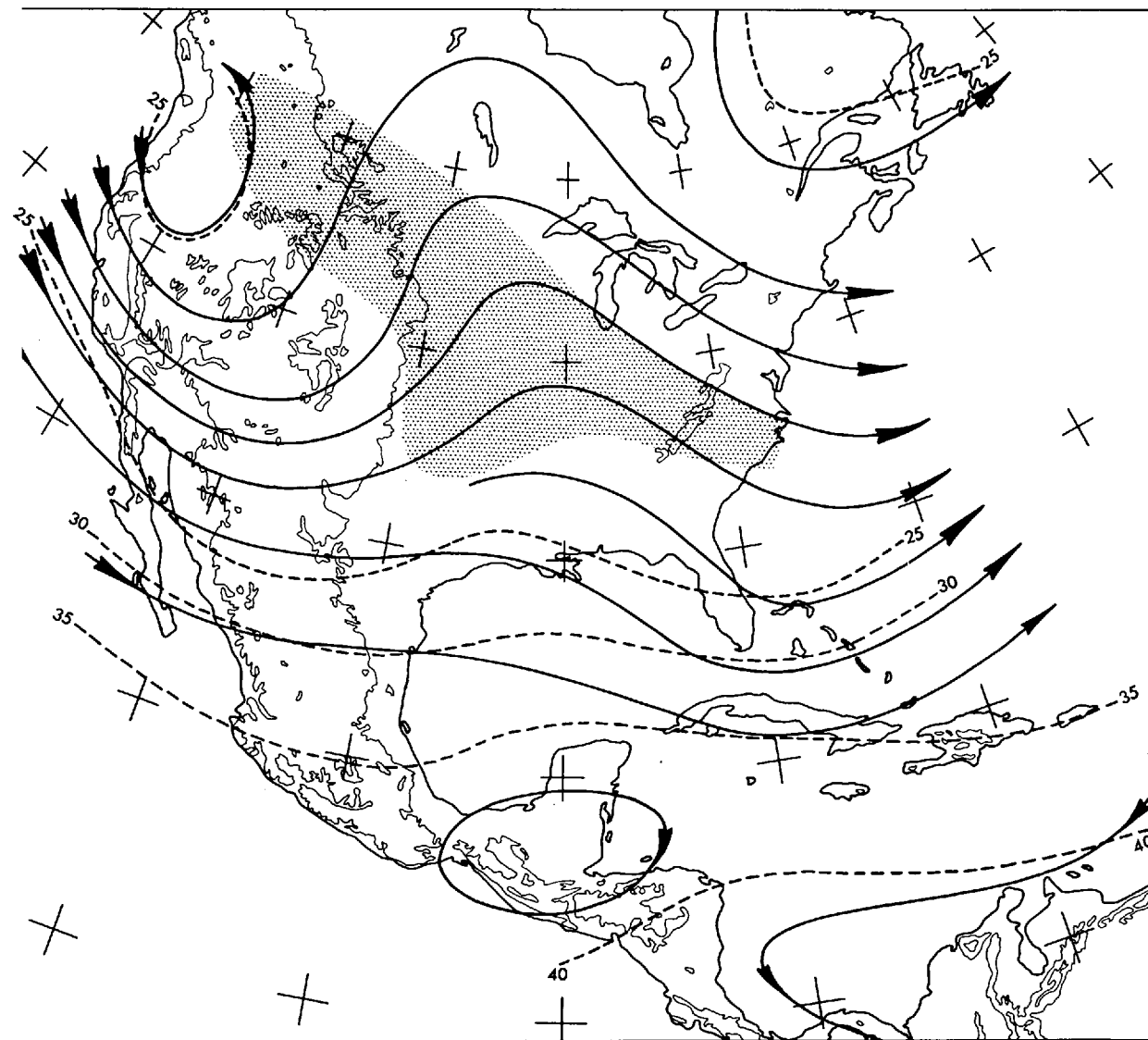


Fig.50. Isentropic relative-flow chart for potential temperature 60°C , 4 May 1961. The pecked lines are isobars, labelled in tens of mb, and the stippled area shows where there is extensive high cloud.

Fig.51

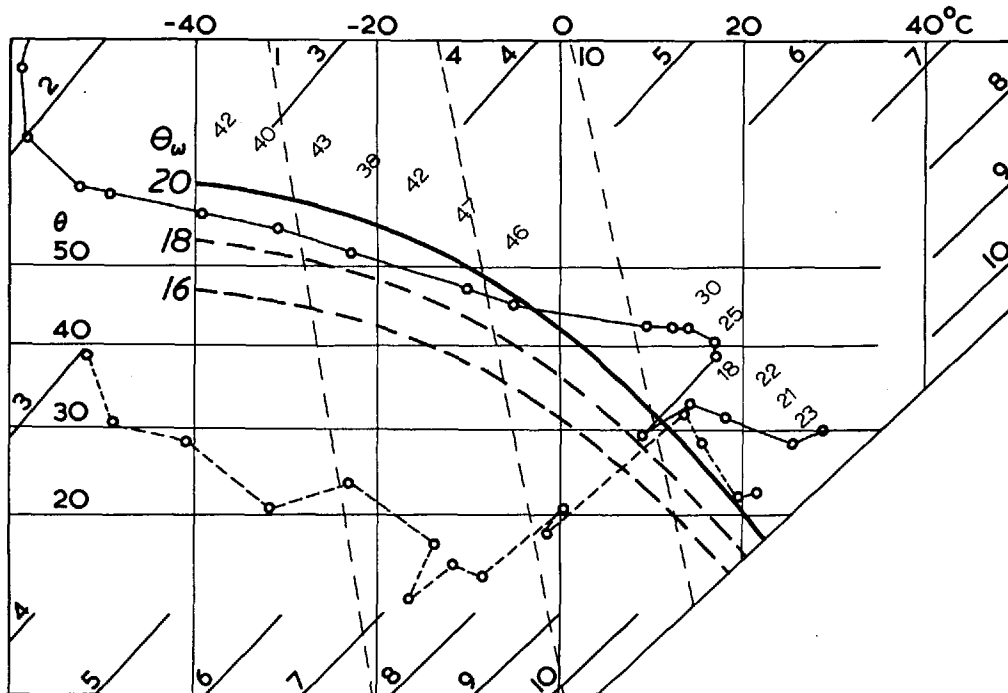


Fig.51. Sounding for San Antonio (72253; marked 'S' on Fig.48), 2330 G.M.T., 4 May 1961. Wind speeds in kts are entered on the left of the sounding. Mexican air is present between about 750 and 550 mb.

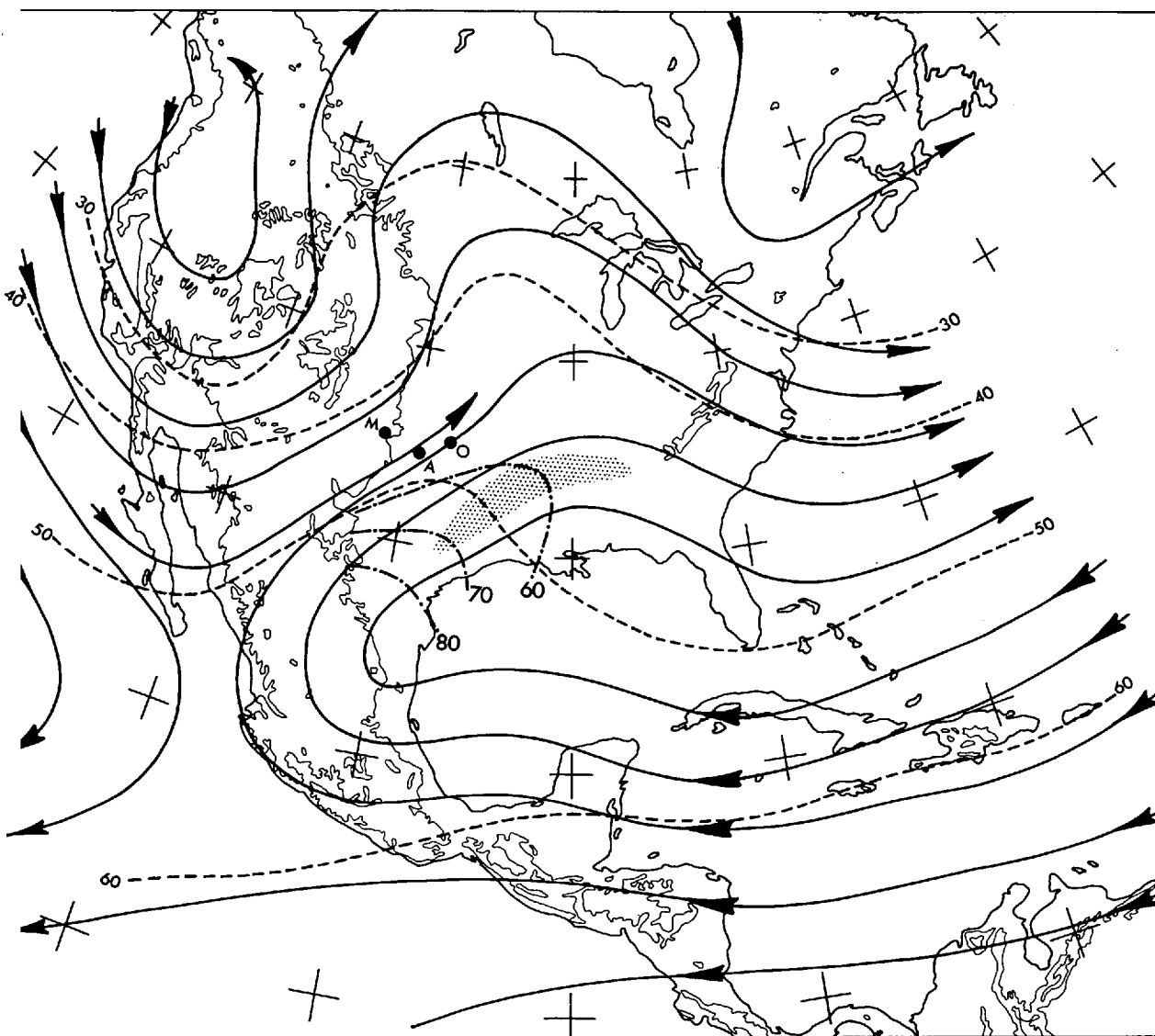


Fig.52. Isentropic relative-flow chart for potential temperature 46°C , 4 May 1961. The pecked lines are isobars at 100 mb intervals, labelled in tens of mb. The dot-dash lines over Texas are isobars also labelled in tens of mb, which show the approximate level of the base of the Mexican plume. The stippled area shows where small amounts of altocumulus and low cirrus cloud appeared within the flow. As on Fig.48, 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.

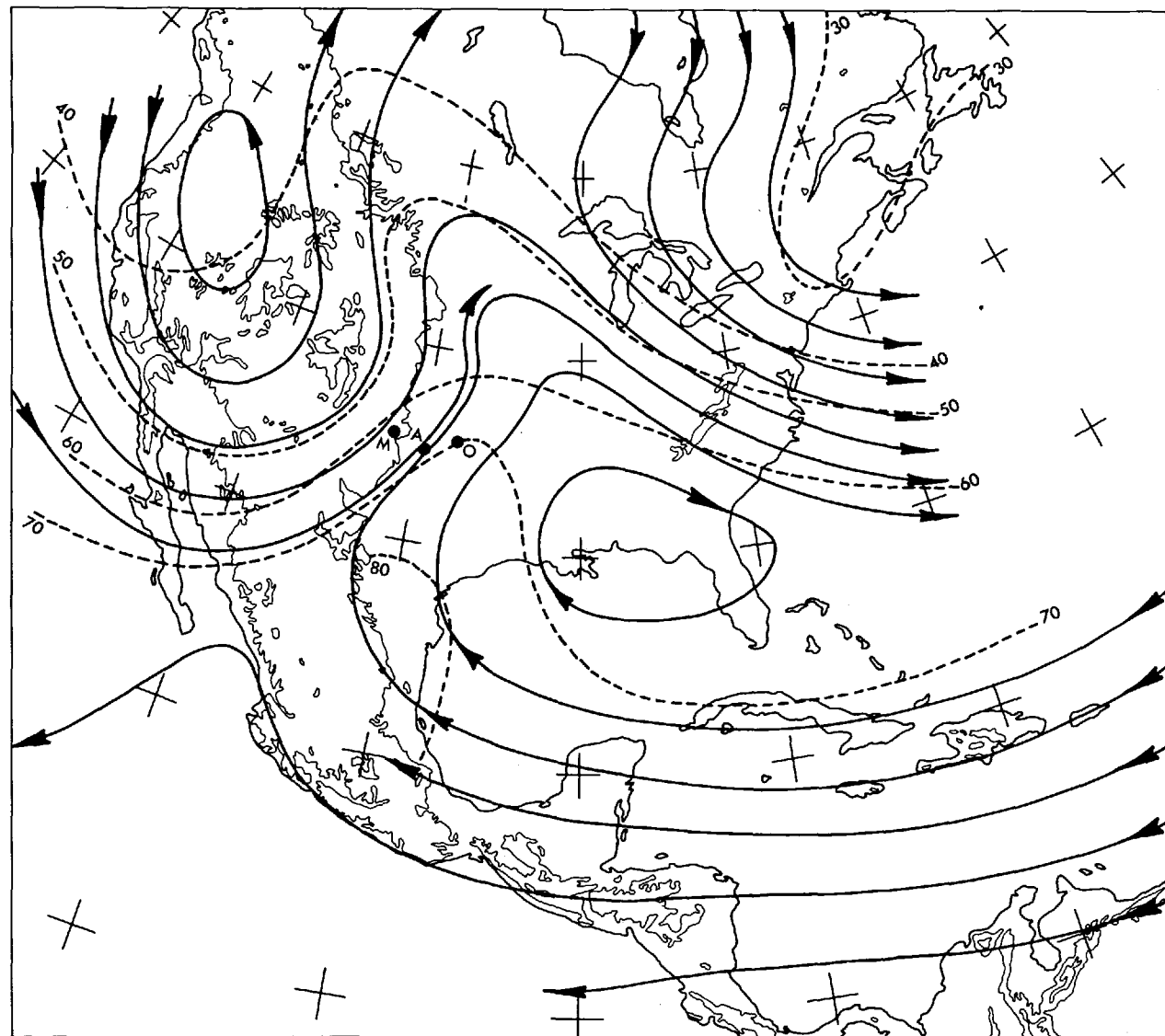


Fig.53. Isentropic relative-flow
chart for potential
temperature 37°C, 2330
G.M.T., 4 May 1961.
Details as in Fig.52.
The limiting streamline
of continental tropical
air is drawn through A.

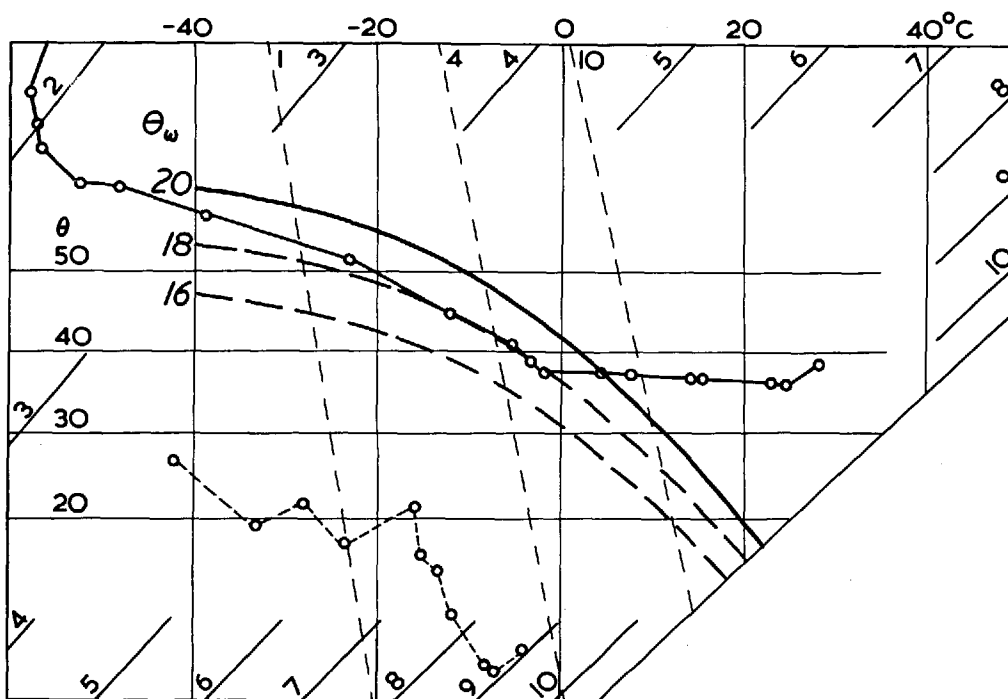


Fig. 54. Sounding for Amarillo (72363; marked 'M' on isentropic charts) 2330 G.M.T., 4 May 1961. The adiabatic layer ($\Theta = 37^{\circ}\text{C}$) below 620 mb is the continental tropical air which has been produced previously by heating mainly over arid northwestern Mexico.

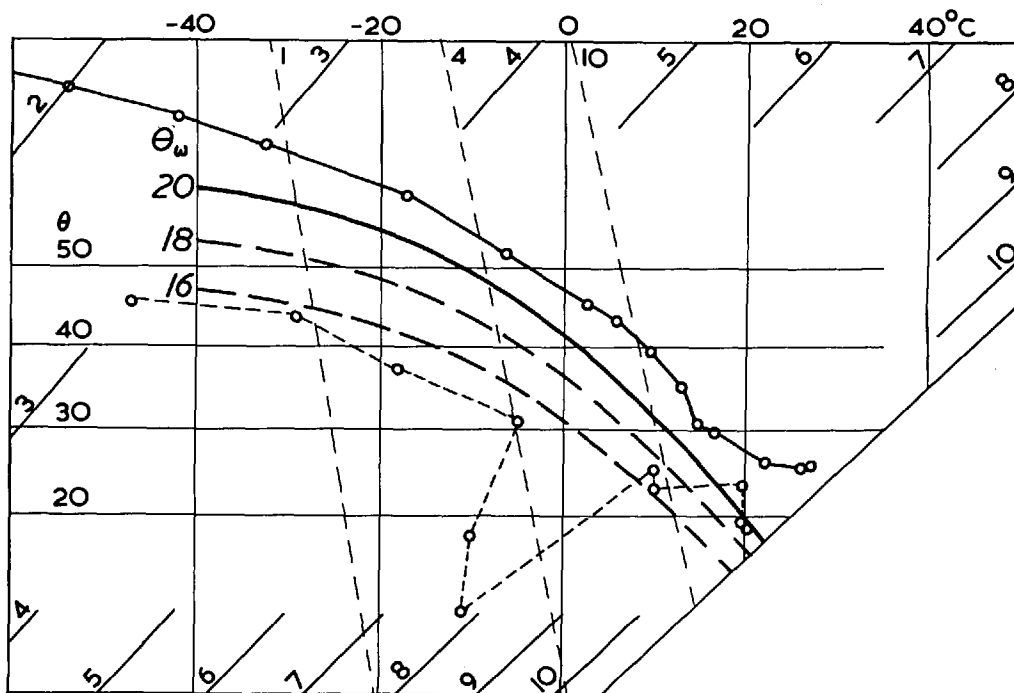


Fig. 55. Sounding for Swan Island (78501; marked 'SI' on Fig. 48), 2330 G.M.T., 4 May 1961. The trade wind moist layer is present up to 820 mb.

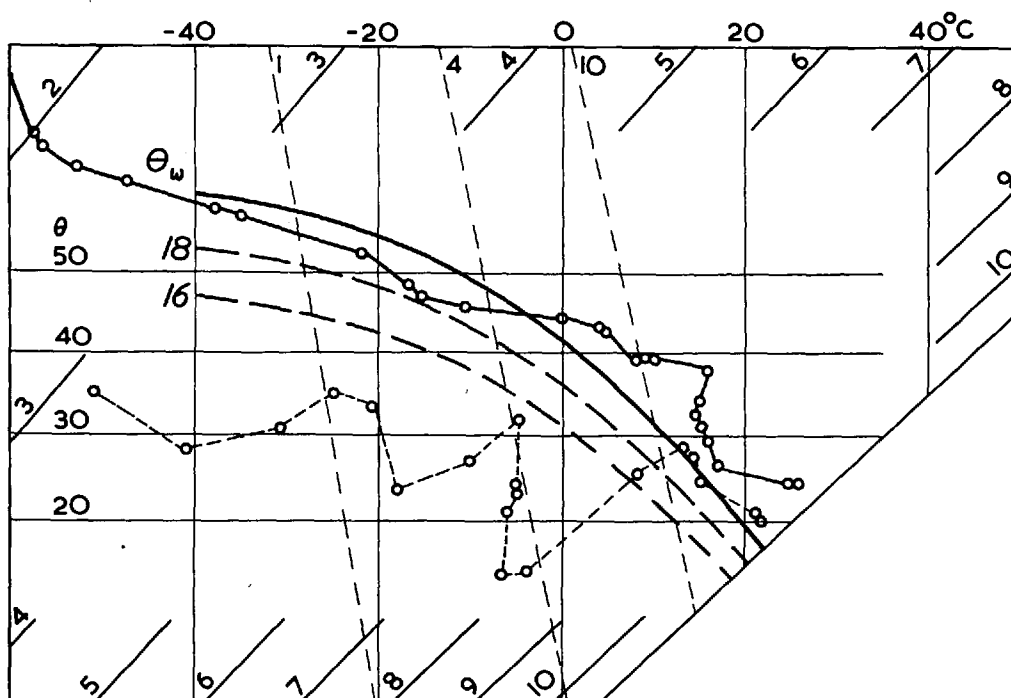


Fig. 56. Sounding for Houston (72243; marked 'H' on Fig. 48), 2330 G.M.T., 4 May 1961. The moist layer is present below 820 mb and between 790 and 490 mb is the Mexican plume.

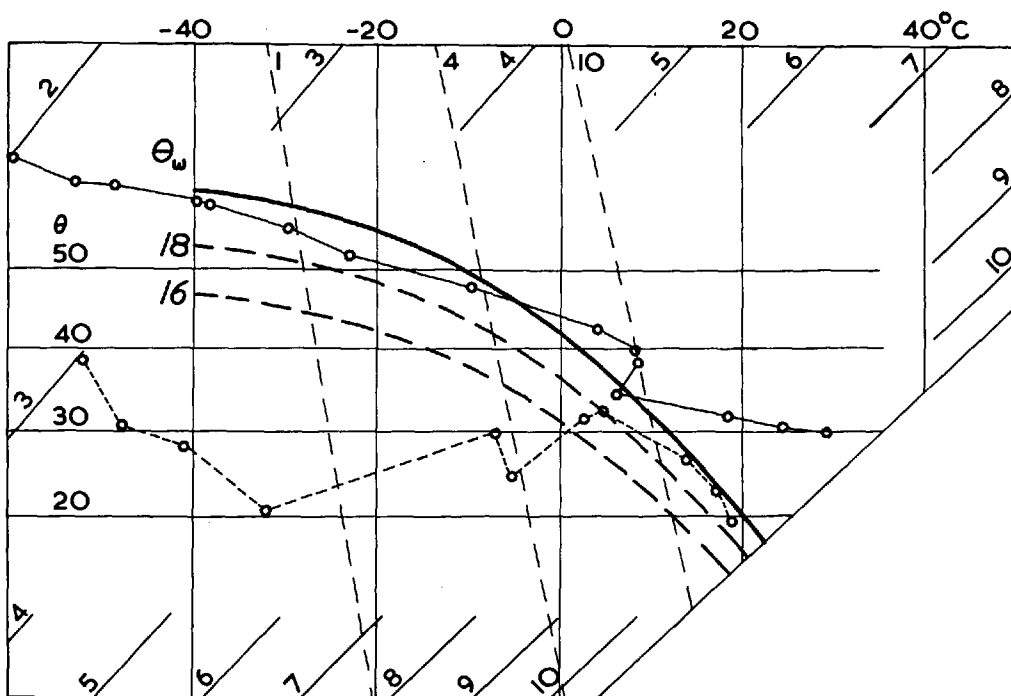


Fig. 57. Sounding for Fort Worth (72259; marked 'F' on Fig. 48), 2330 G.M.T., 4 May 1961.

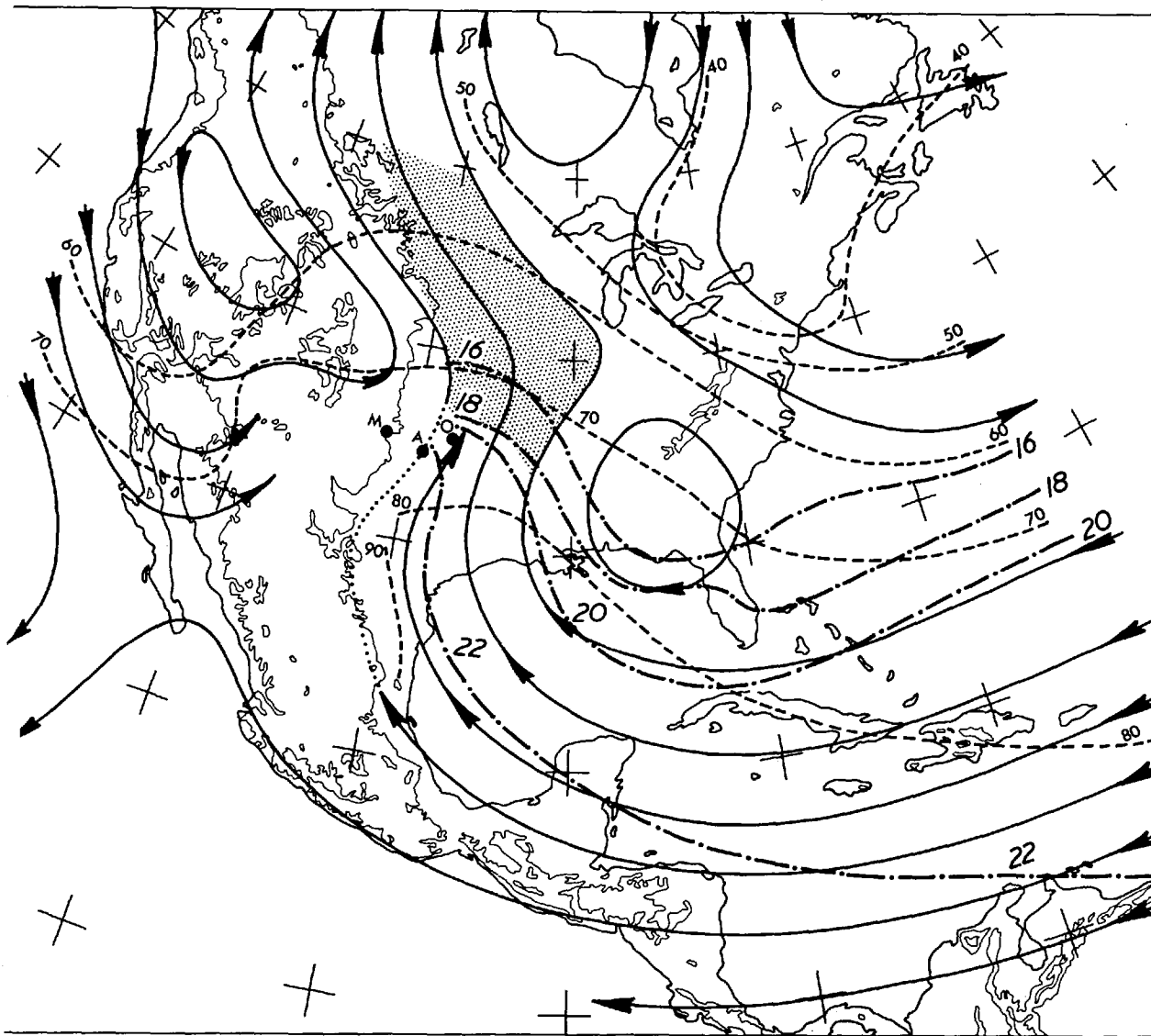


Fig. 58. Isentropic relative-flow chart for potential temperature 32°C , 2330 G.M.T. 4 May 1961. Details as in Fig. 52, with the addition of isopleths (dot-dash lines labelled in $^{\circ}\text{C}$ in italic figures) showing the average wet-bulb potential temperature in the lowest 100 mb. The stipple shows a belt of medium-level cloud extending away northward from the general thunderstorm area. The limiting streamline, coinciding with the surface position of the dry line, is shown as a dotted line where heating has produced an adiabatic layer with potential temperature somewhat in excess of 32°C .

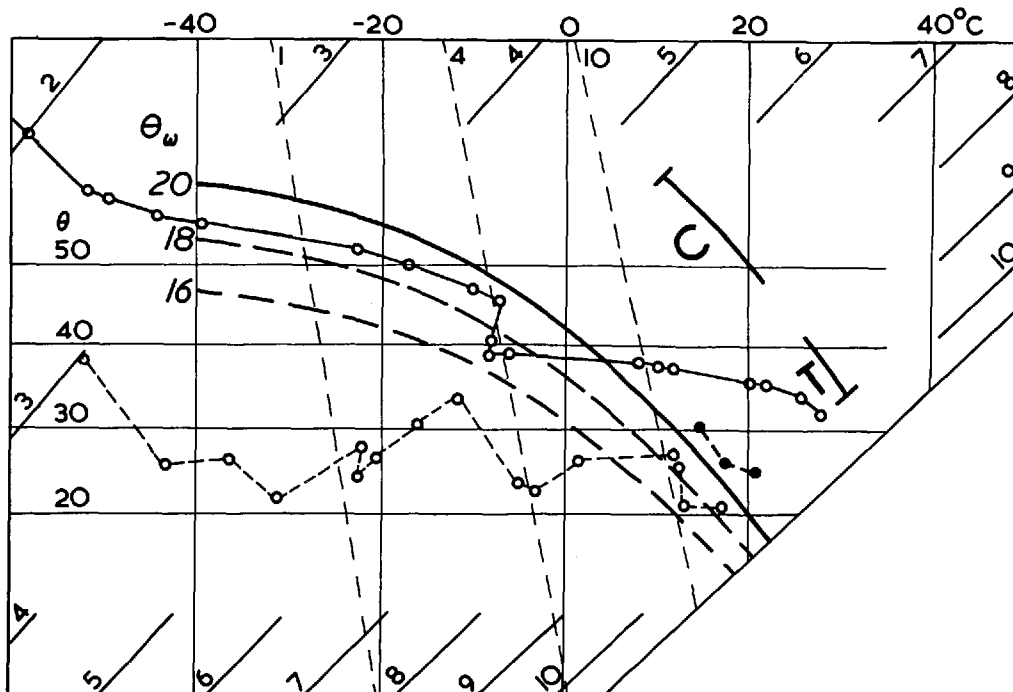


Fig. 59. Sounding for Altus (72352; marked 'A' on Figs. 48, 52, 53 and 58), 2330 G.M.T., 4 May 1961. The distribution of the wet-bulb potential temperature in the moist trade wind layer (T) is shown by full circles; above is the layer of 'continental tropical' air (C) with the characteristic potential temperature of about 38°C (see Fig. 54).

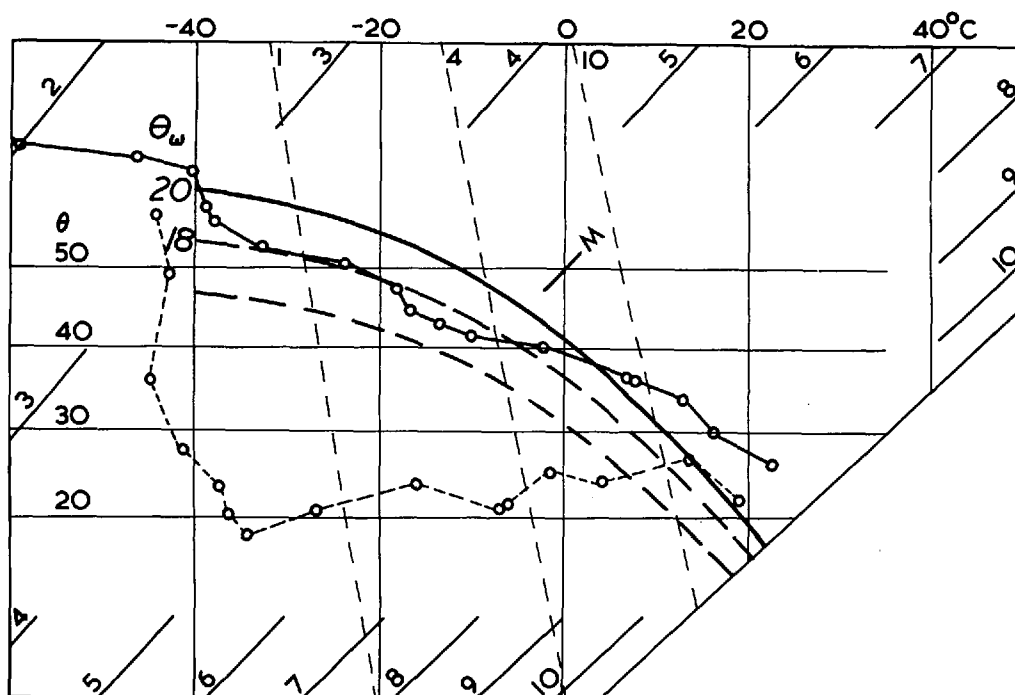


Fig. 60. Sounding for Oklahoma City (72353; marked 0 on Figs. 48, 52, 53 and 58), 2330 G.M.T., 4 May 1961. A shallow layer of air from the Mexican adiabatic layer is probably present at about the level M (550 mb).

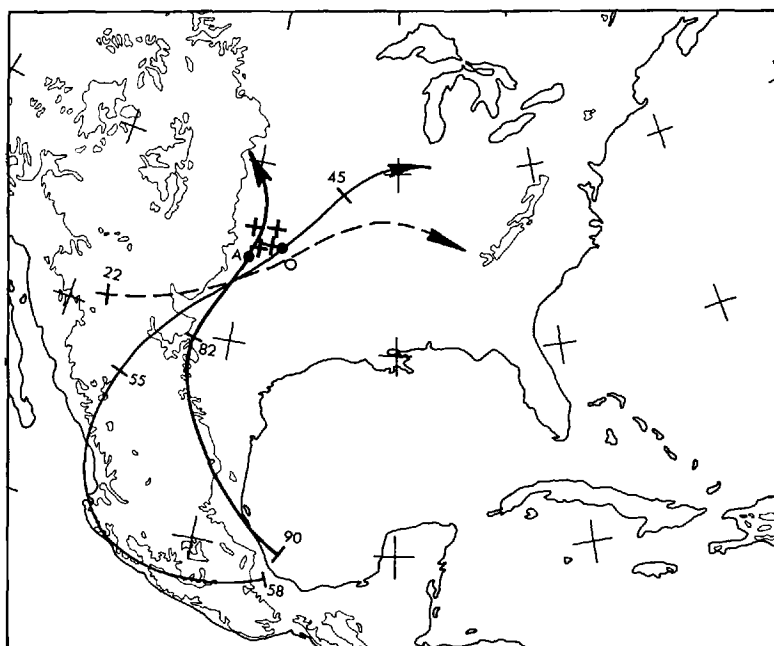


Fig. 61. Composite chart, 4 May 1961. The limiting streamlines are shown for the moist trade wind flow (thick line; height of top of layer indicated in tens of mb) and the top of the Mexican air (thin line; height marked in tens of mb). The pecked line shows the flow in the high troposphere, at about 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).

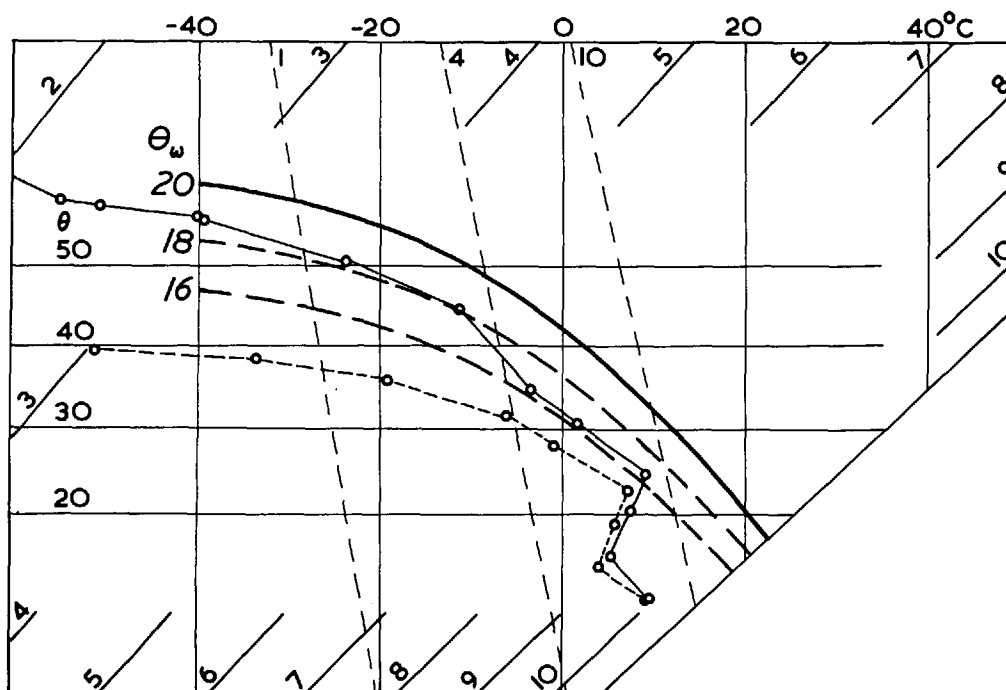


Fig. 62. Sounding for Topeka (72456; marked T on Fig. 48), 2330 G.M.T., 4 May 1961. An almost saturated layer which contains altocumulus is indicated between 830 and 630 mb, with the wet-bulb potential temperature to be anticipated in the thunderstorm downdraughts.

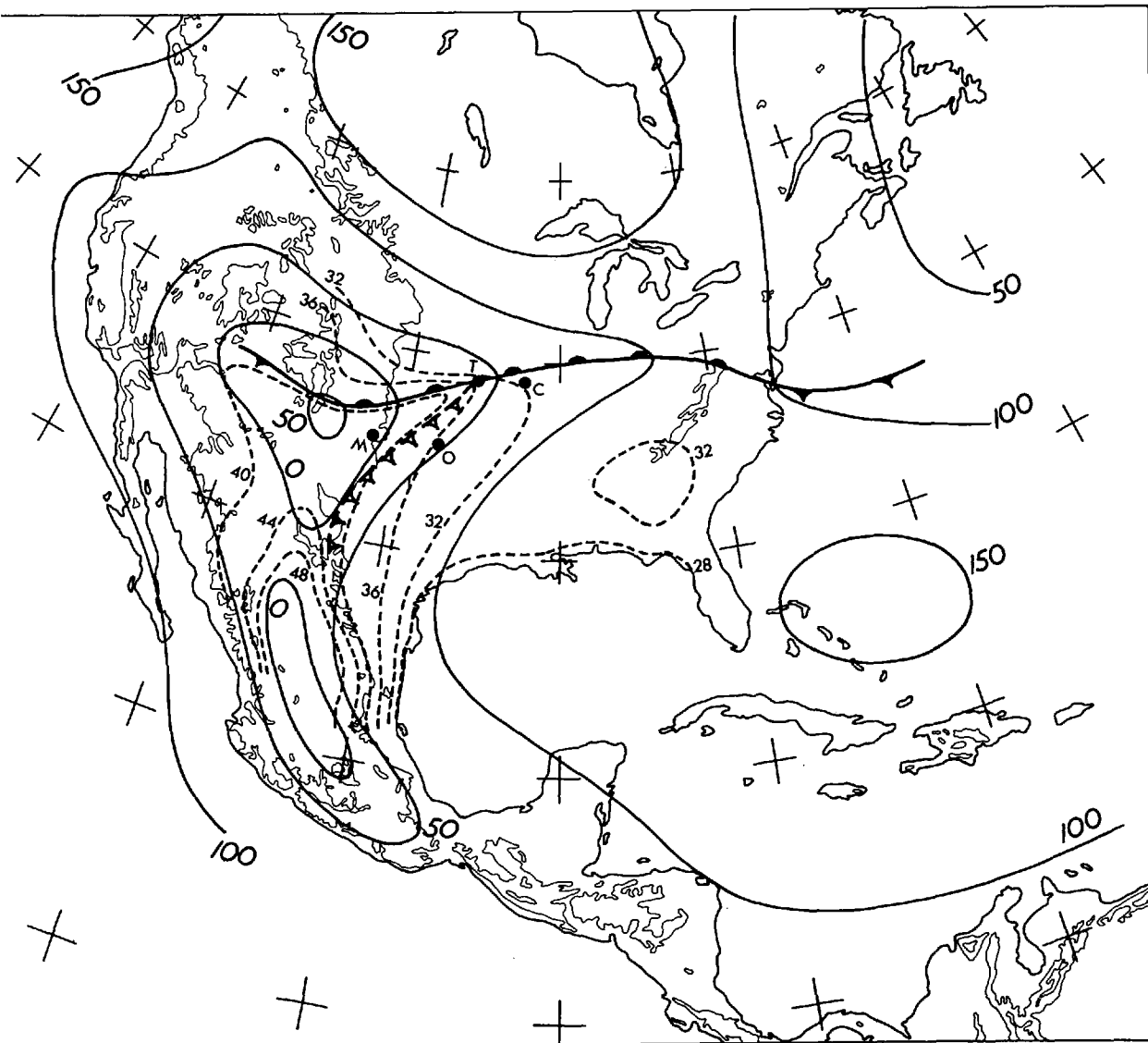


Fig. 63. Contours of the 1000 mb surface (continuous lines labelled in m), 2330 G.M.T., 26 May 1961, and conventional frontal analysis (dry-line indicated by pecked line with open cold front symbols). Also included are isopleths of screen level potential temperature (approximately the day's maximum) at intervals of 4°C . The places marked by full circles are Amarillo (M), Oklahoma City (O), Topeka (T) and Columbia (C).

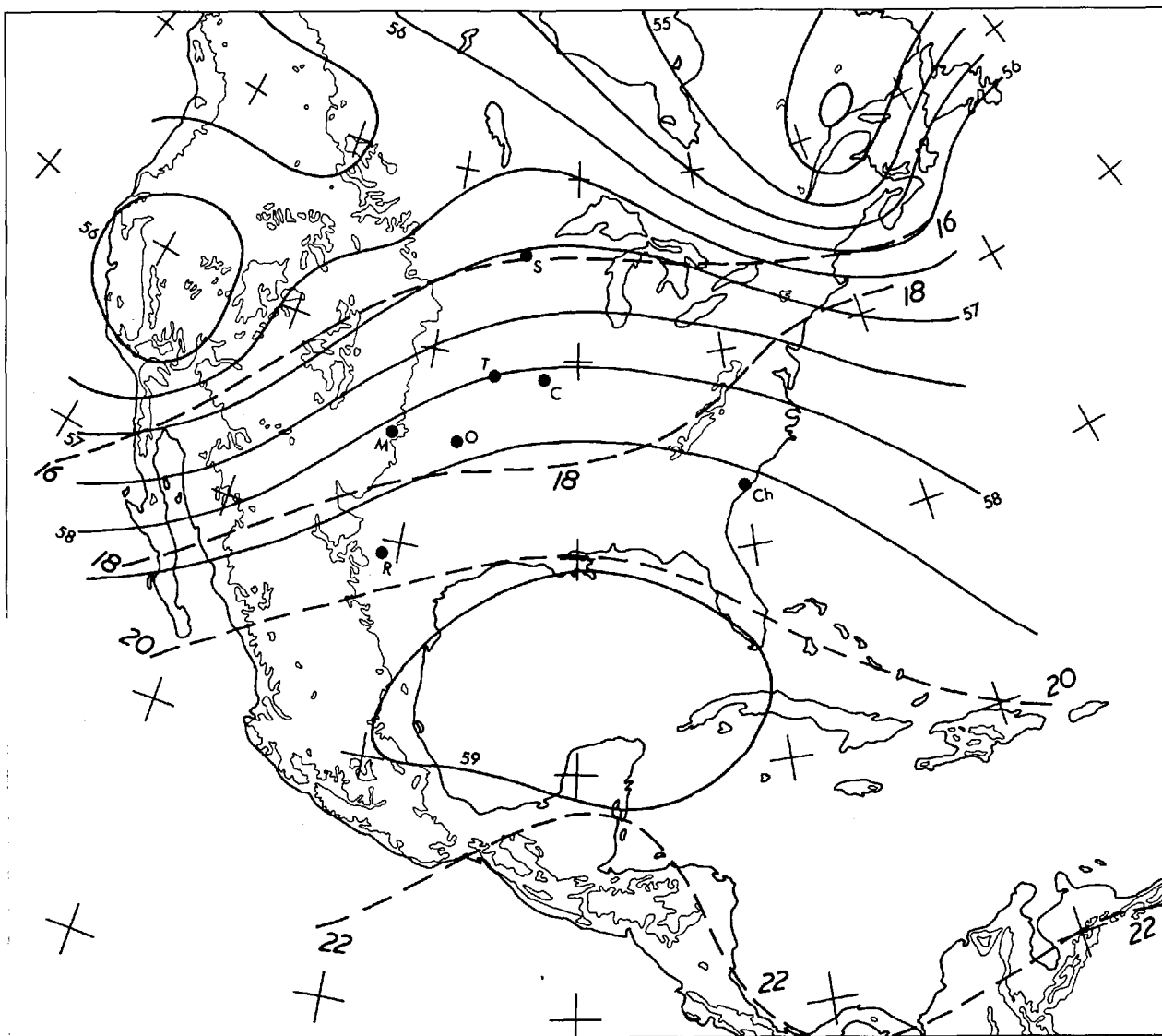


Fig. 64. Contours of the 500 mb surface (continuous lines labelled in hundreds of m) and isopleths of Θ_s at 500 mb (pecked lines labelled in $^{\circ}\text{C}$), 2330 G.M.T., 26 May 1962. The places marked by full circles are, from west to east, Amarillo (M), Del Rio (R), Oklahoma City (O), Topeka (T), St. Cloud (S), Columbia (C), and Charleston (Ch).

Fig. 65

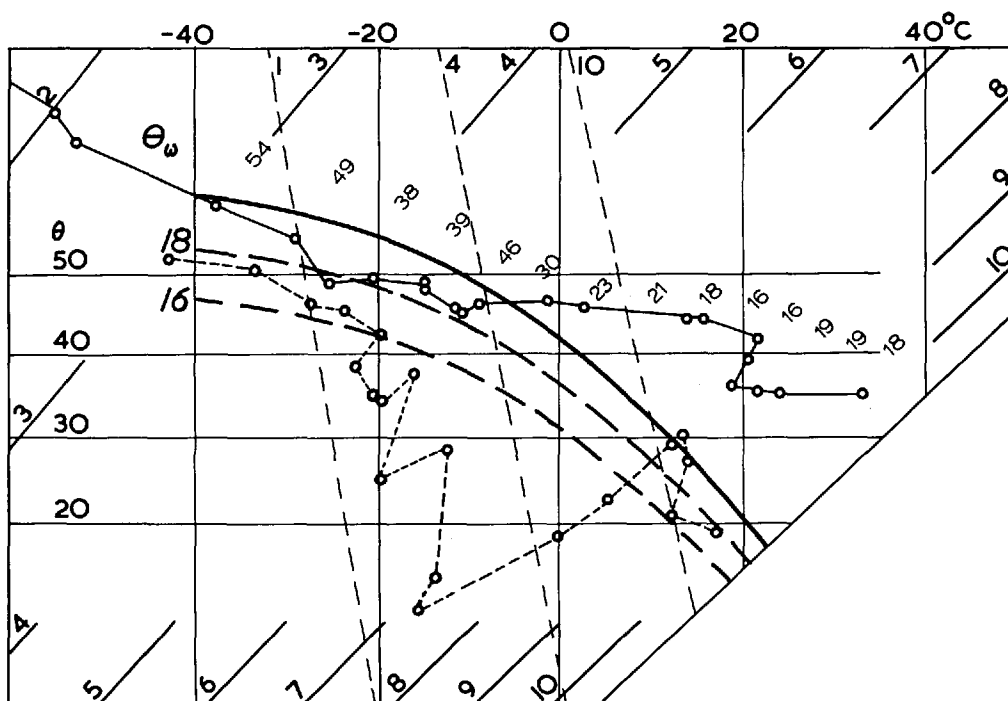


Fig. 65. Sounding for Del Rio (72261; marked R on Fig. 64), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered to the right of the sounding. Mexican air is present between about 790 and 510 mb.

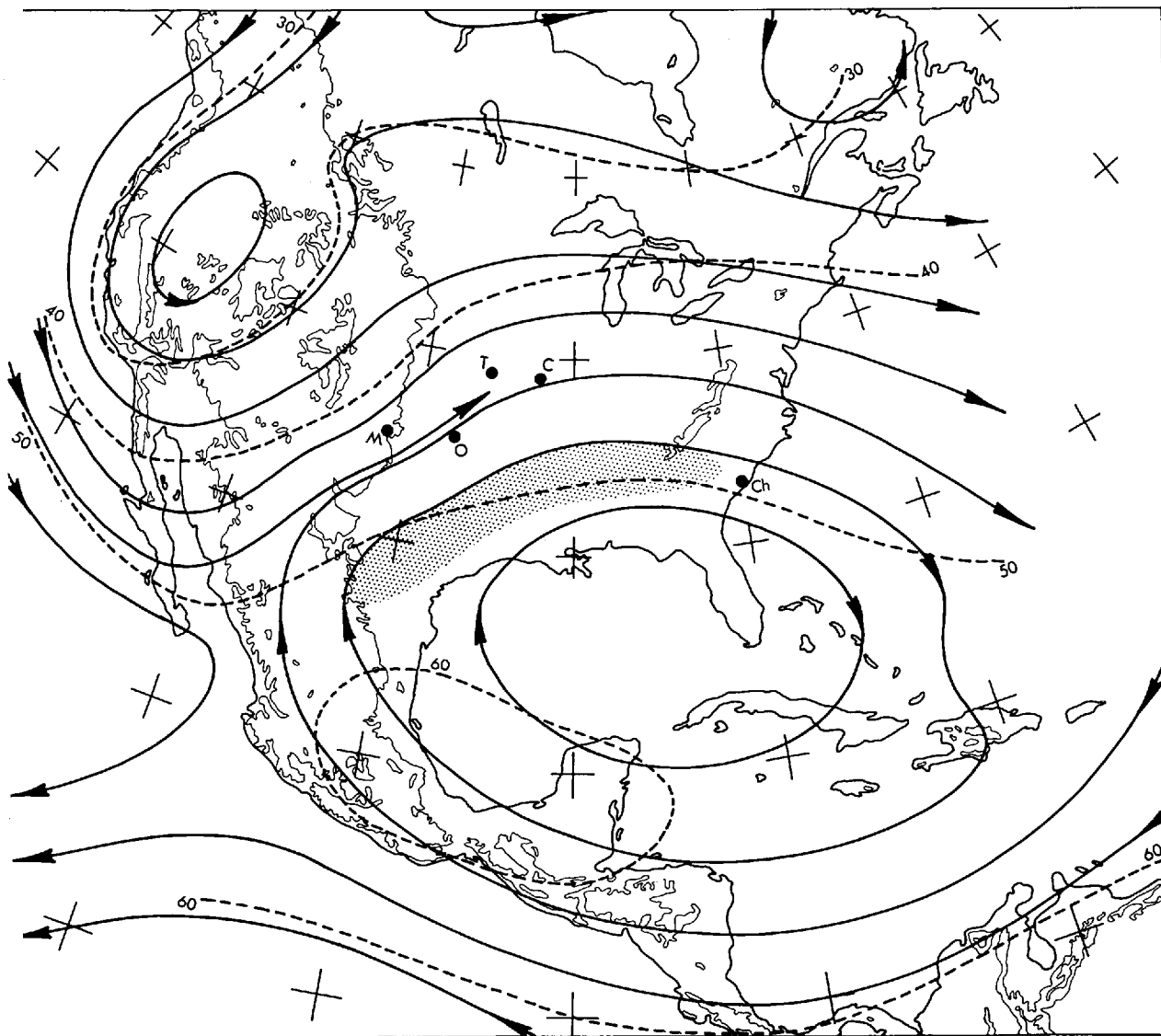


Fig.66. Isentropic relative-flow chart for potential temperature 47°C, 2330 G.M.T., 26 May 1962. Height contours (pecked lines) are labelled in tens of mb. Stippling marks a belt of extensive altocumulus and low cirrus. The limiting streamline of the Mexican air passes close to Oklahoma City (O) and Columbia (C), and east of Amarillo (M) and Topeka (T). The position of Charleston is shown by the letters Ch.

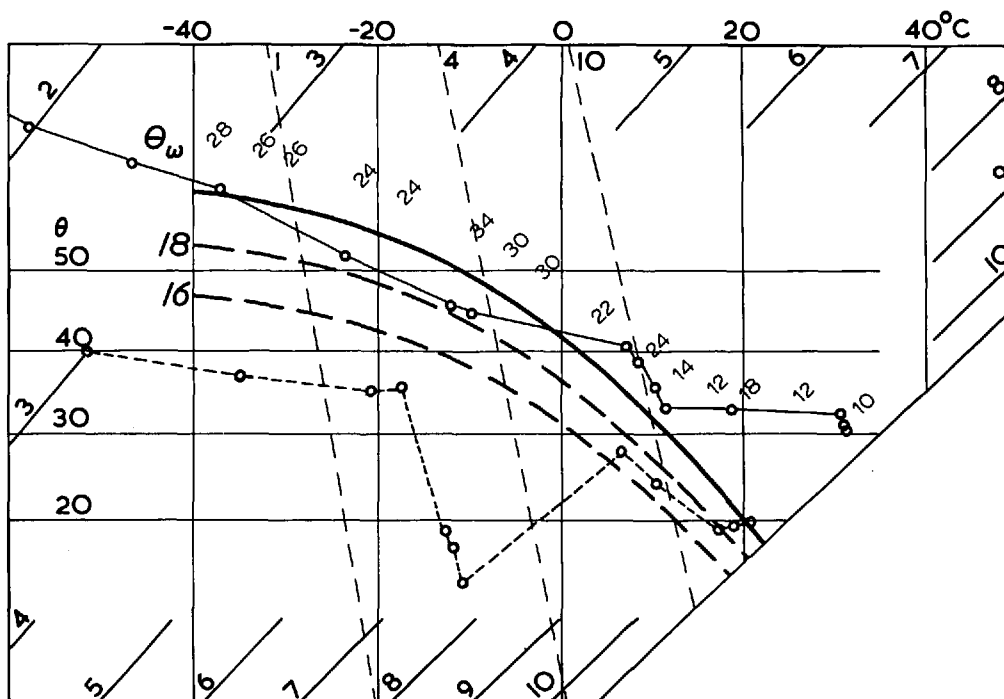


Fig. 67. Sounding for Charleston (72208; marked Ch on Fig. 64), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. Mexican air is present between about 680 and 520 mb.

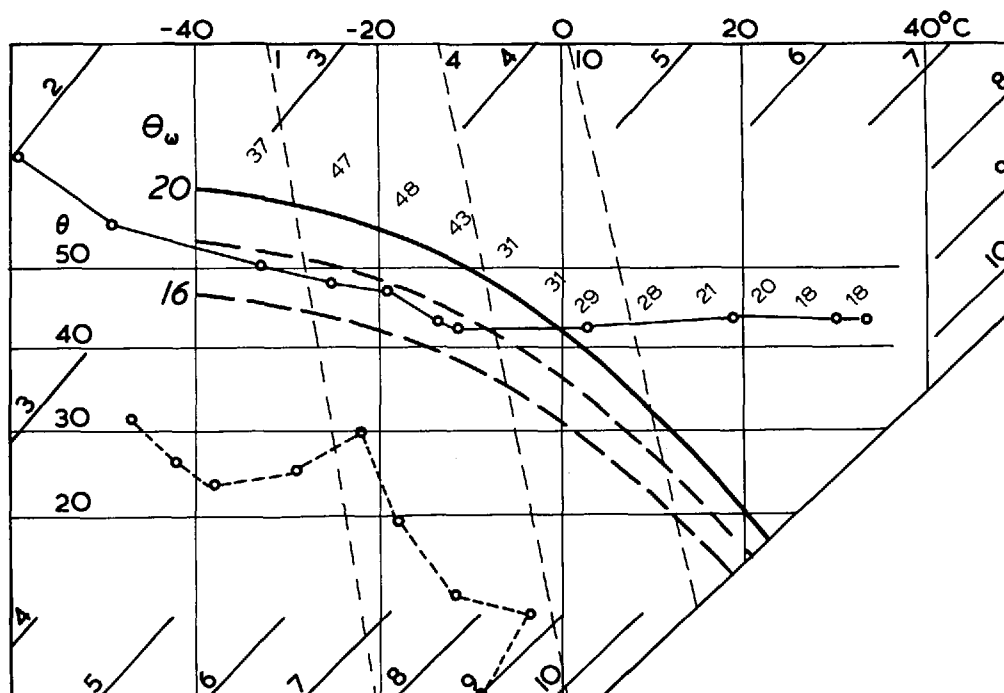


Fig. 68. Sounding for Amarillo (72363; marked M on Figs 63, 64 and 66), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. An adiabatic layer of continental tropical air is present below 520 mb.

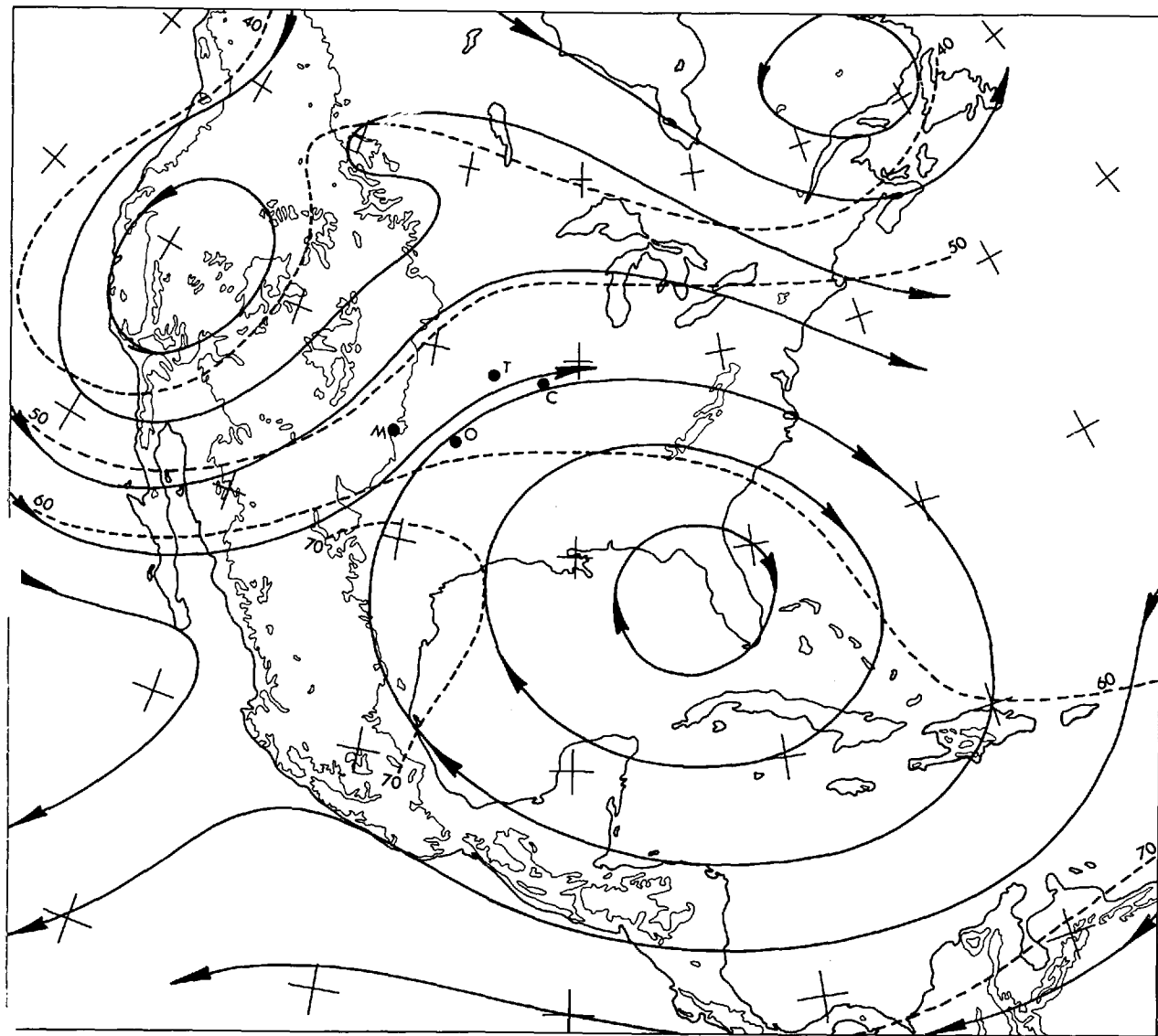


Fig. 69. Isentropic relative-flow chart for potential temperature $42\frac{1}{2}^{\circ}\text{C}$, 2330 G.M.T., 26 May 1962. Details as on Fig. 66. The limiting streamline in the continental tropical air passes just to the south of Columbia (C).

Fig.70

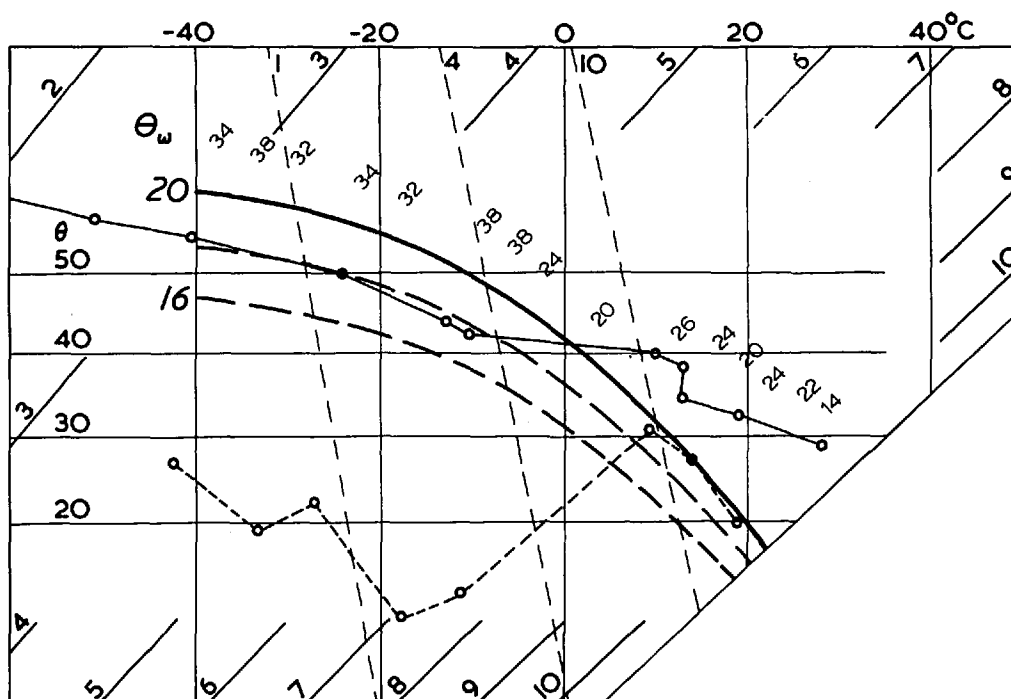


Fig. 70. Sounding for Columbia (72445; marked C on Figs. 63, 64, 66 and 69), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. Continental tropical air is present between 740 and 520 mb, and the moist layer, $\Theta_w = 22^\circ\text{C}$, in the bottom 200 mb.

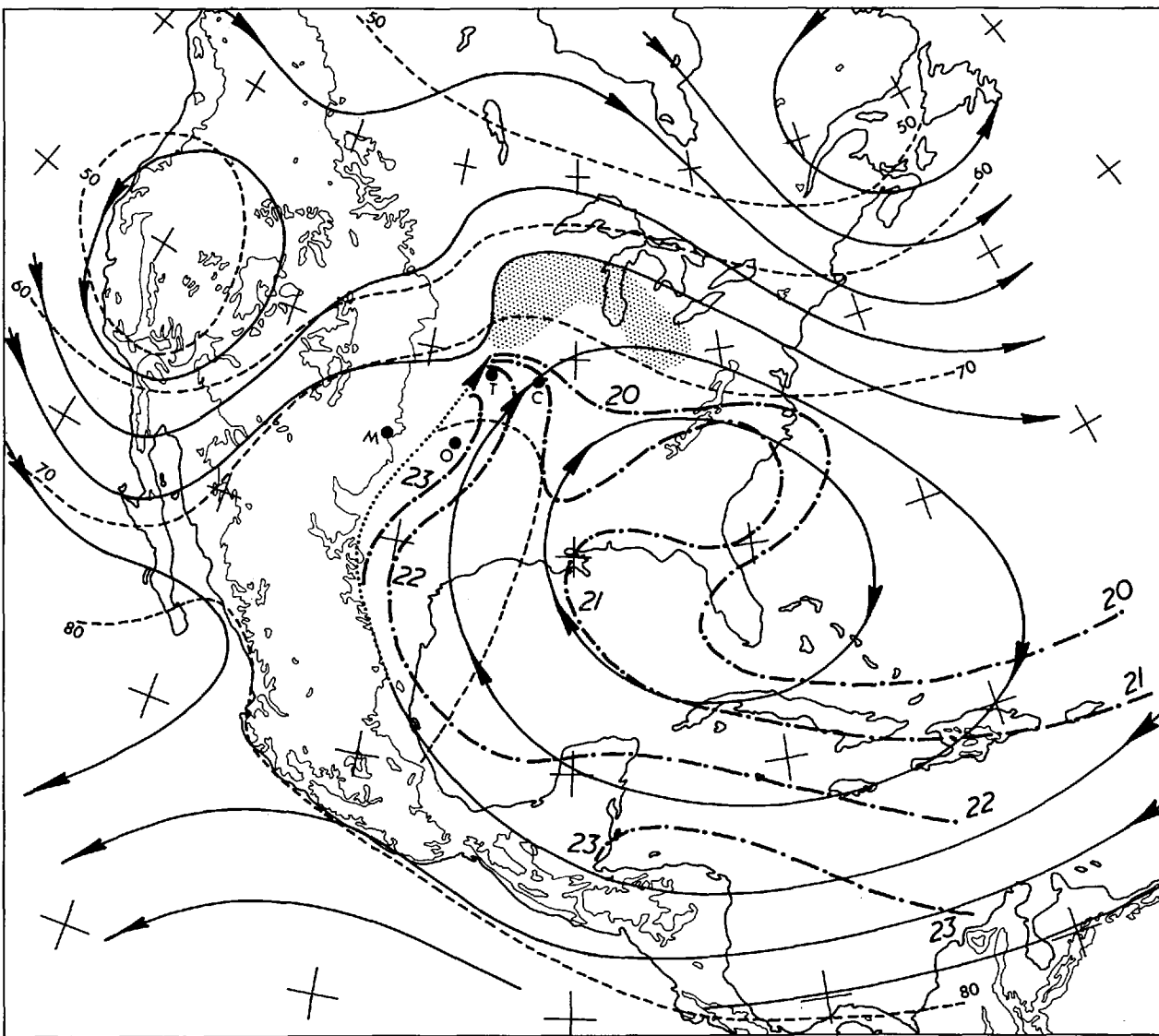


Fig. 71. Isentropic relative-flow chart for potential temperature 35°C , 2330 G.M.T., 26 May 1962. Height contours at intervals of 100 mb (pecked lines) are labelled in tens of mb. The stippled area represents a belt of middle-level cloud formed in air from the thunderstorm downdraughts after ascent over the warm front. The dot-dash lines labelled in $^{\circ}\text{C}$ show the distribution of mean wet-bulb potential in the lowest 100 mb of the trade-wind 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).

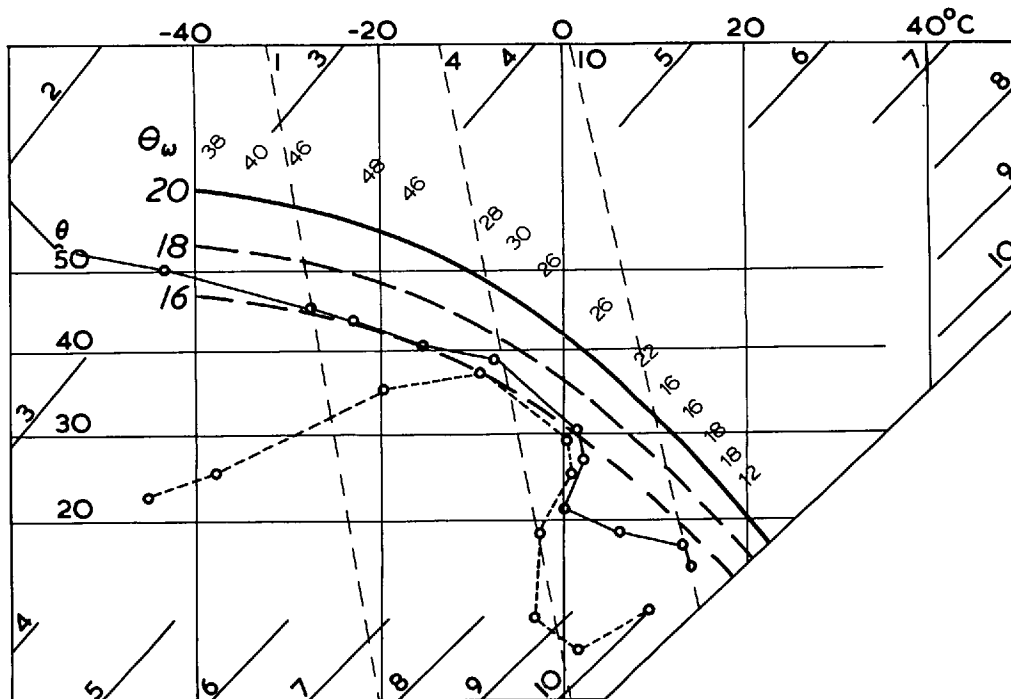


Fig. 72. Sounding for St. Cloud (72855, marked S on Fig. 64), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. Cloud is present in the layer of practically saturated air between about 700 and 580 mb, which has ascended over the warm front after descending in thunderstorm down-draughts. This sounding closely resembles that from Topeka on 4 May 1961 (Fig. 62).

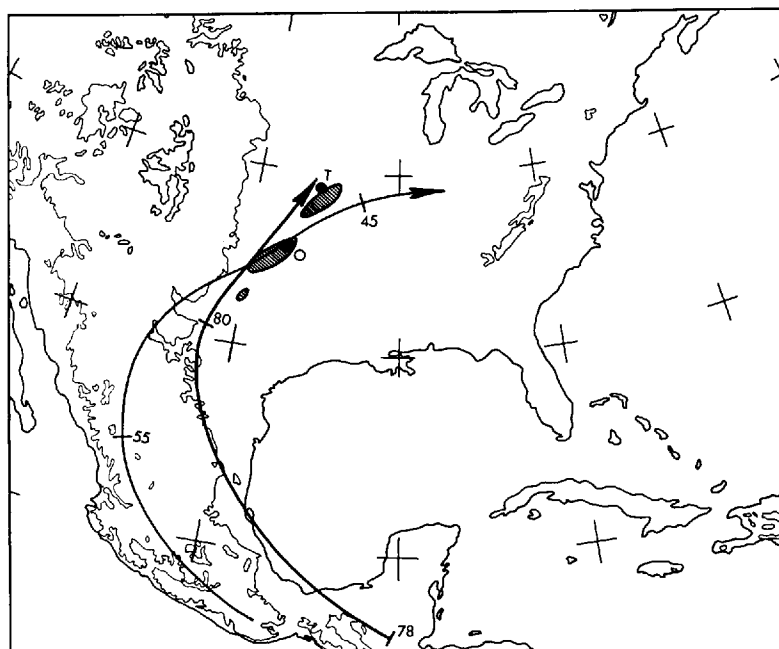


Fig. 73. Composite chart, 26 May 1962. The limiting streamlines are shown for the moist trade-wind flow (thick line; height of top of layer indicated in tens of mb) and the top of the Mexican air (thin line; height marked in tens of mb). Also marked are the areas of severe storms (hatched areas) and the positions of Oklahoma City (O) and Topeka (T).

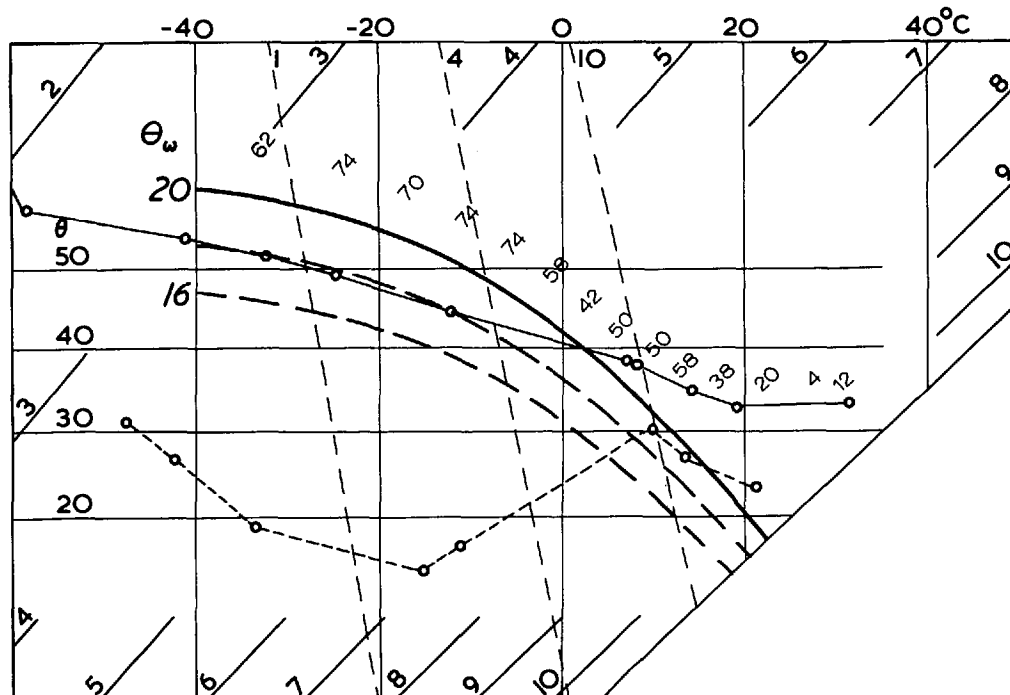


Fig. 74. Sounding for Topeka (72456; marked T in Figs. 64, 71 and 73), 2330 G.M.T., 26 May 1962. Wind speeds in kts are entered on the right of the sounding. The unstable stratification strongly resembles that at Altus on 4 May 1961 (Fig. 59).

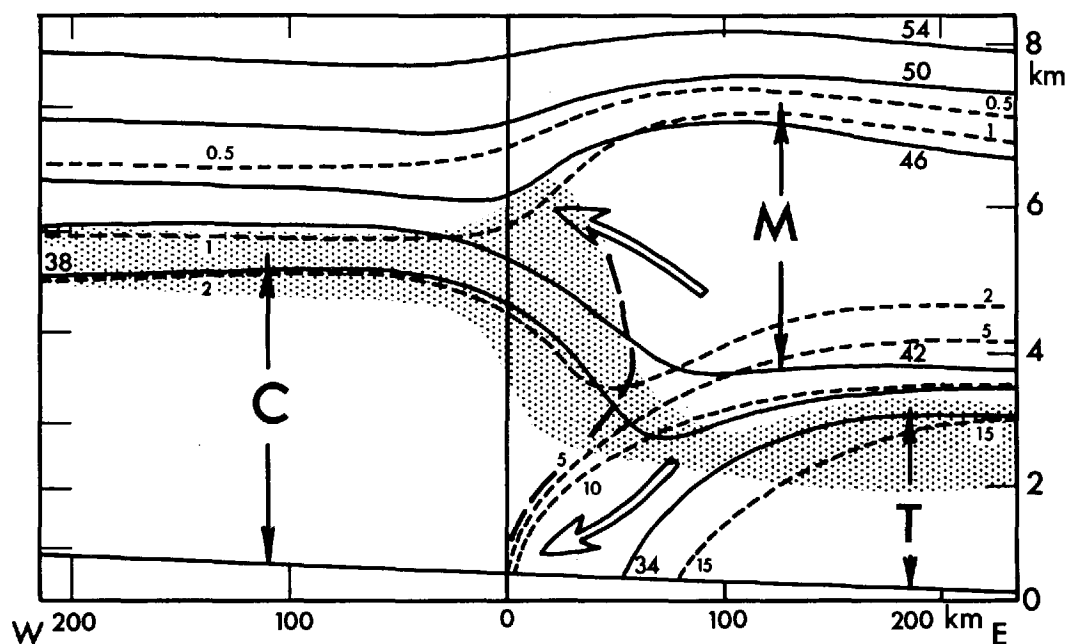


Fig. 75. Schematic cross-section of a typical dry-line, showing as a function of distance in km from its surface position the distribution of potential temperature (isopleths at 4°C intervals), and mixing-ratio (pecked lines labelled 0.5, 1, 2, 5, 10 and 15 g/kg). The Continental Tropical air is denoted 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.

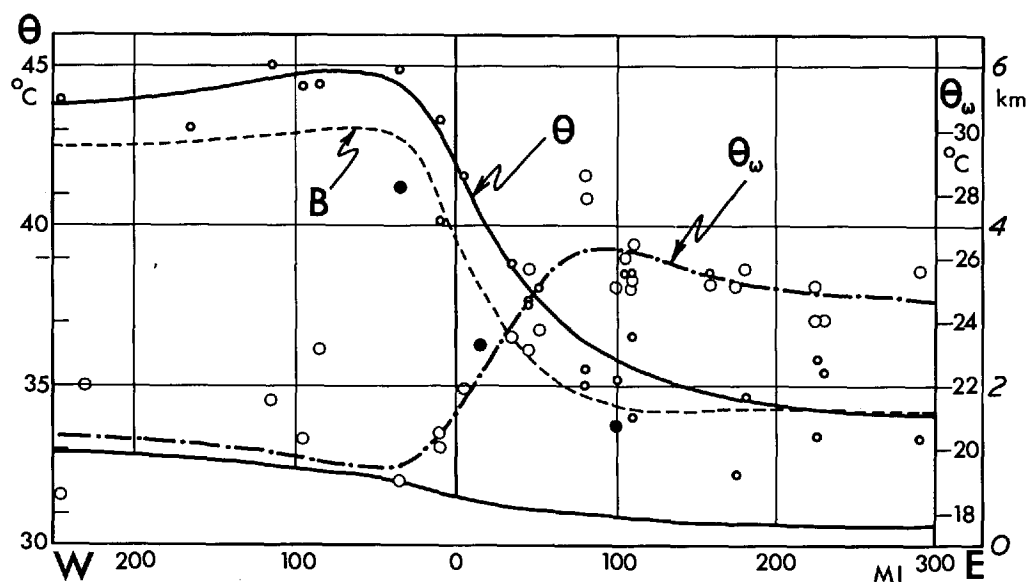


Fig. 76. Distribution of potential temperature (small circles, continuous line), wet-bulb potential temperature (large circles, dot-dash line) and adiabatic condensation level (pecked line marked B), of screen-level air, 2330 G.M.T., 26 May 1962. The observations are entered according to distance from the dry-line. The full circles mark the level of cumulus base estimated from photographs taken by a reconnaissance aircraft west of Oklahoma City. The gradients of properties close to the dry-line may have been much stronger than indicated in the figure, based on smoothed synoptic data.

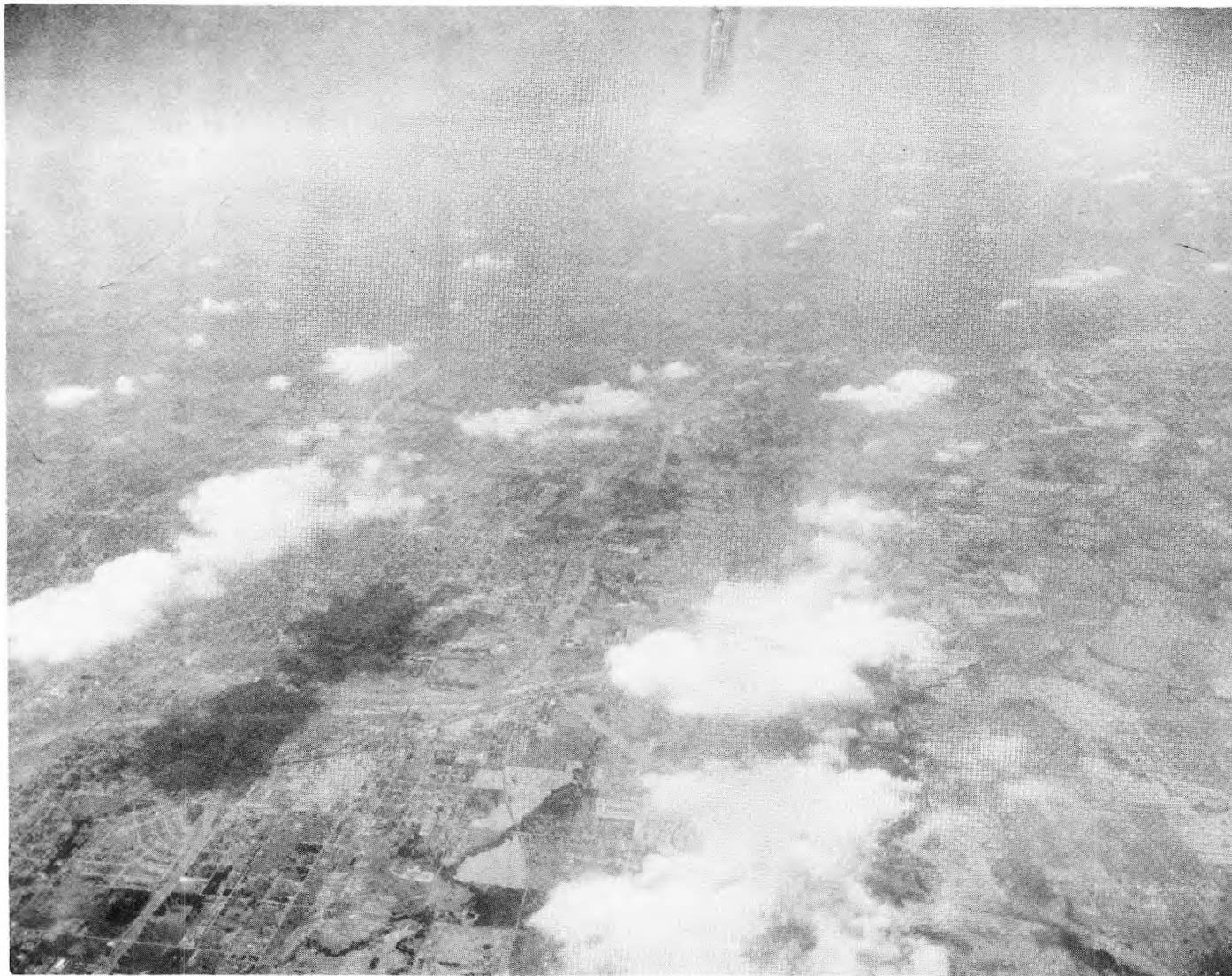


Fig.77. View northwards from Oklahoma City (from position '1' on Fig.78) at 14,000 ft, 1954 G.M.T. (1354 C.S.T.), 26 May 1962. There are scattered small cumulus over and east of the City, disappearing farther west towards the dry-line.

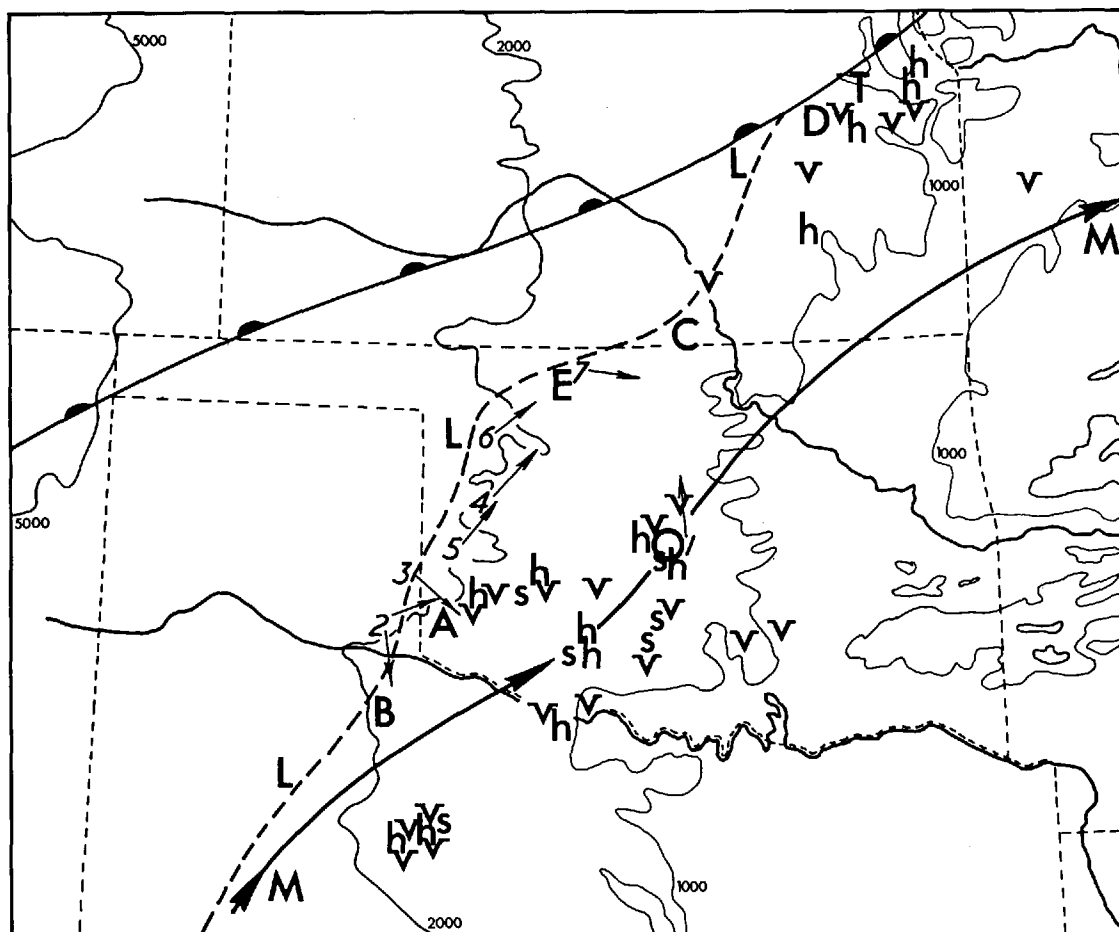


Fig. 78. Map showing position of the warm front, the dry-line (LL), the limiting streamline (MM) of the Mexican air, and of Oklahoma City (O, in right centre) and of Topeka (T, top right). Ground contours are drawn for heights of 1000, 2000 and 5000 ft above sea level, and the state borders and principal 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 6 and the associated arrows show the positions and directions of view of the pictures in Figs. 77, 79-86. The letters A to E show the positions of cumulonimbus groups mentioned in the text and figure legends.

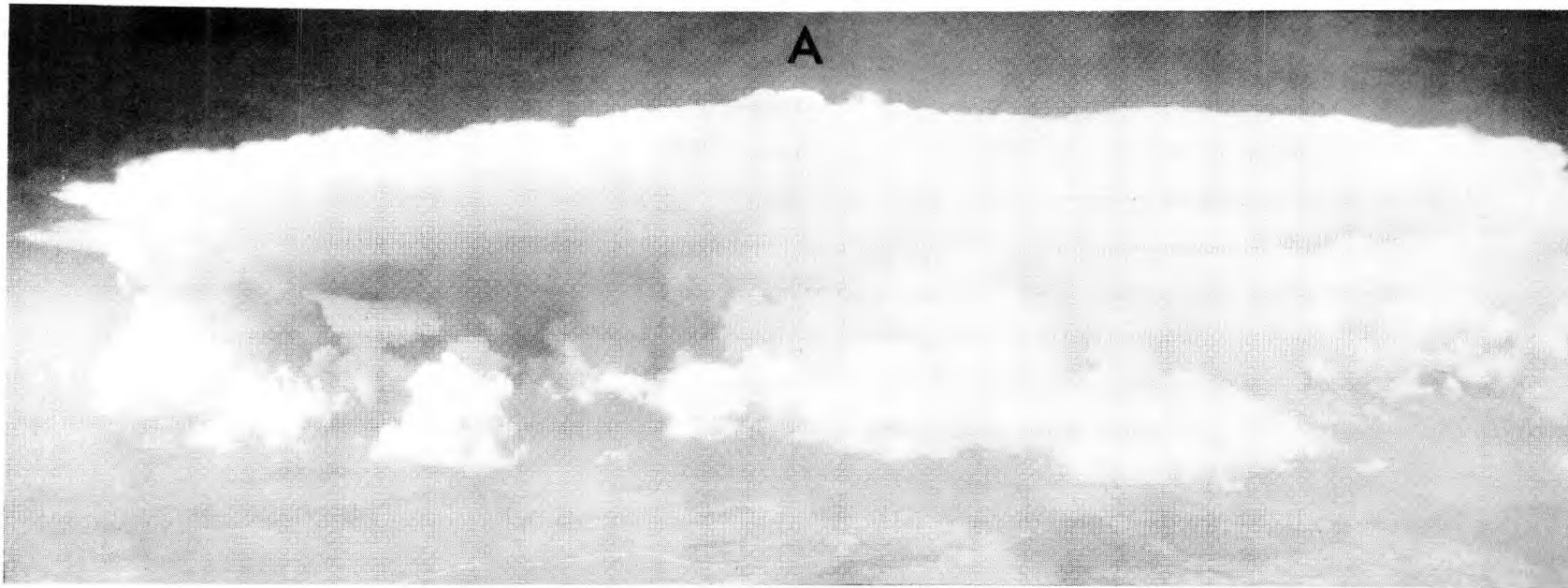


Fig. 79. View looking ENE of mature cumulonimbus (A of Fig. 78), from about 25,000 ft near dry-line (position 2, Fig. 78), 2100 G.M.T. (1500 C.S.T.), 26 May 1962.

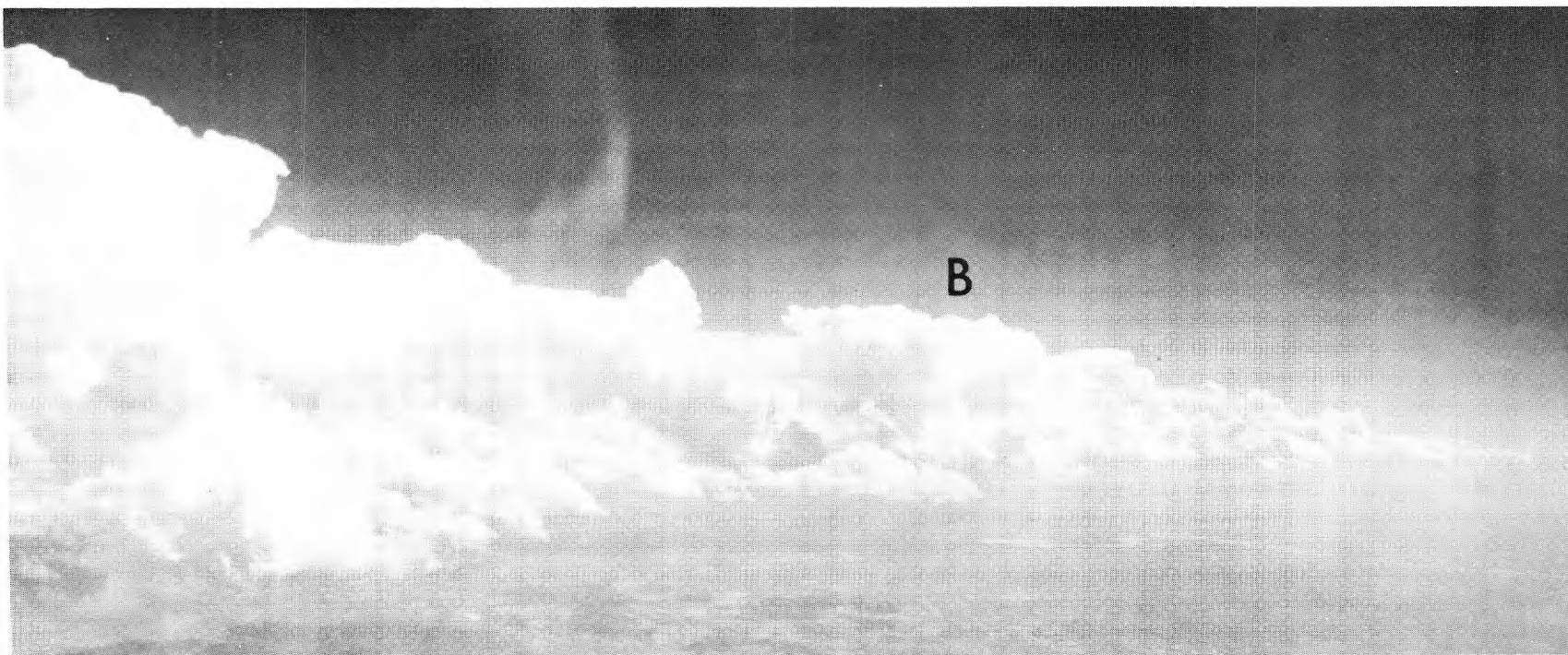


Fig. 80. View looking S from 25,000 ft near dry-line (position 2, Fig. 78), 1435 C.S.T., 28 May 1962. There is intense cumulus and cumulonimbus development just east of the dry-line. The group of cumulonimbus beginning to develop anvils in the right centre of the picture are marked B in Fig. 78.

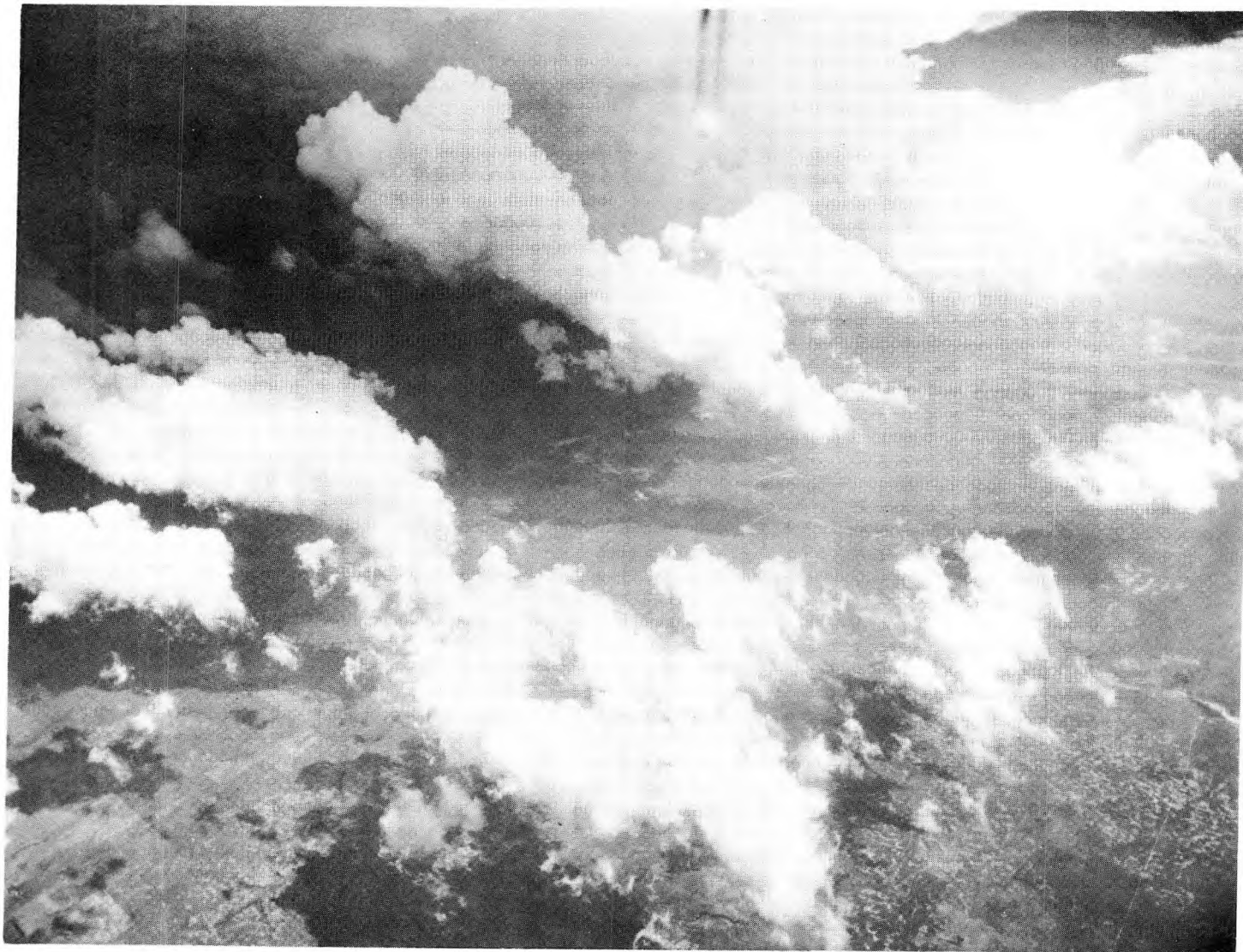


Fig.81. View looking SE from 25,000 ft at the dry-line (position 3, Fig.78), 1508 C.S.T., 26 May 1962. The upper parts of the congestus cloud in the left centre, whose lean indicates the marked wind shear, are shaded by the anvils of the cumulonimbus group A.

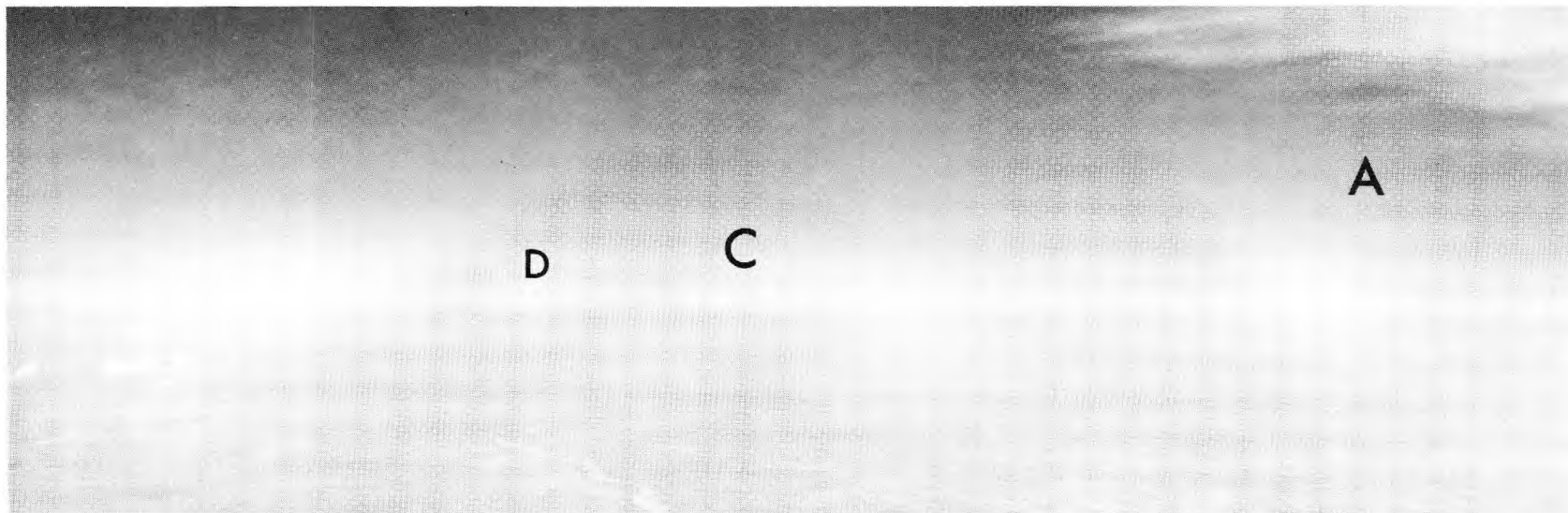


Fig.82. View looking NE from 25,000 ft nearly along the dry-line, 1518 C.S.T., 26 May 1962 (position 4, Fig.78). The recently developed groups of tall congestus clouds in the centre and left centre (faintly seen on the original print and here indicated by the letters C and D) are respectively to develop into the cumulonimbus groups C, about 140 mi away on the Kansas-Oklahoma border, and D, about 280 mi away near Topeka (see Figs. 78, 83, 85 and 86).

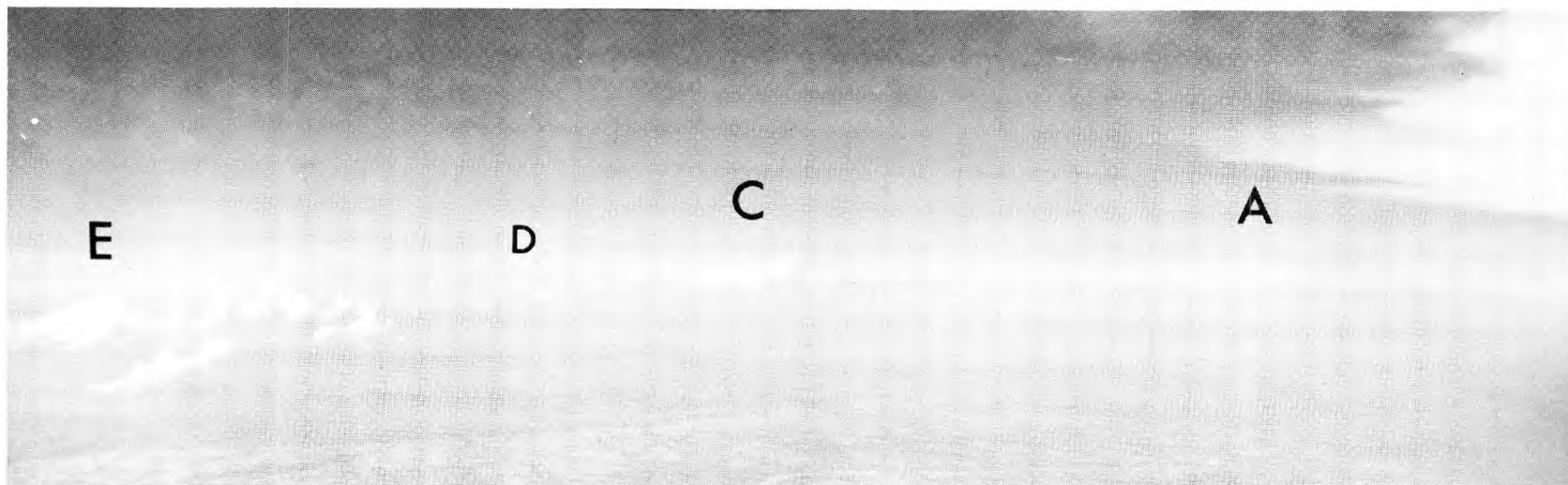


Fig.83. View as in Fig.82, 22 minutes later (1540 C.S.T.), showing clouds of group C commencing anvil formation. Note also the near congestus on the left beginning to tower (later producing cumulonimbus group E), and on the right the edge of the anvil of the cumulonimbus A.

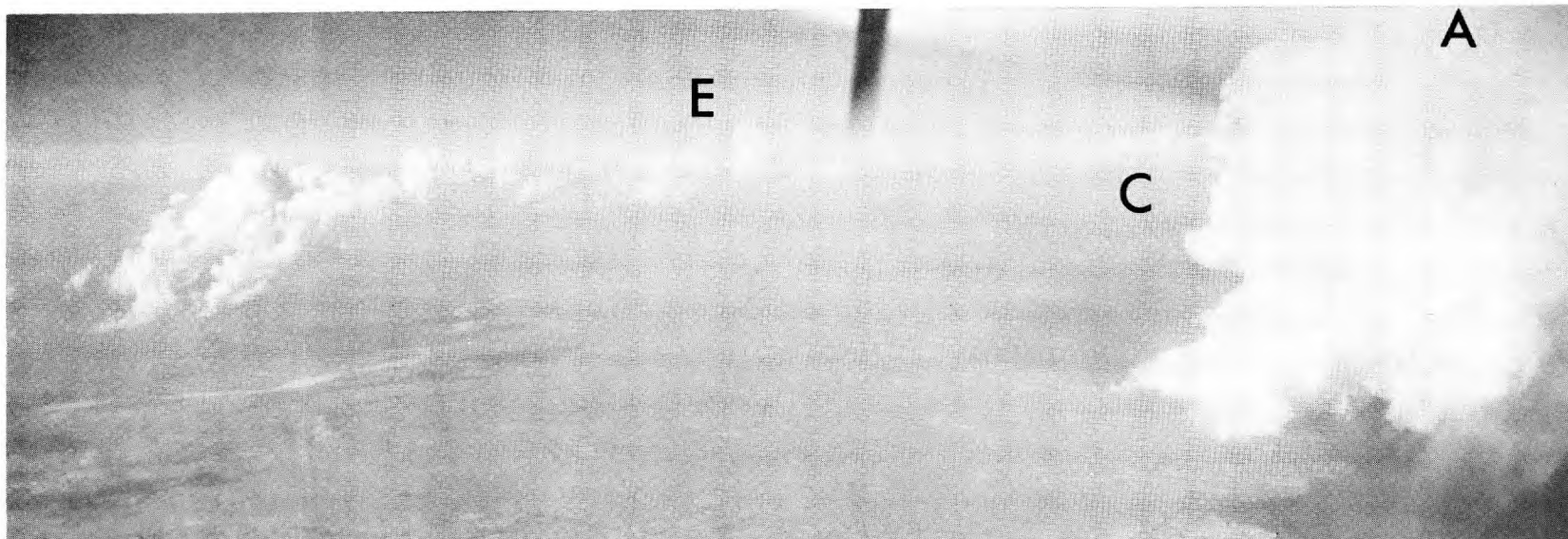


Fig.84. View from 25,000 ft looking NE from position 5, Fig.78, 1602 C.S.T., 26 May 1962. On the right are the flanks and anvil of cumulonimbus group A, partially obscuring the developing anvils of group C. On the left the nearer congestus (E) continue to tower: at this time in three distinct places the tops reach the flight level with precipitation formation.

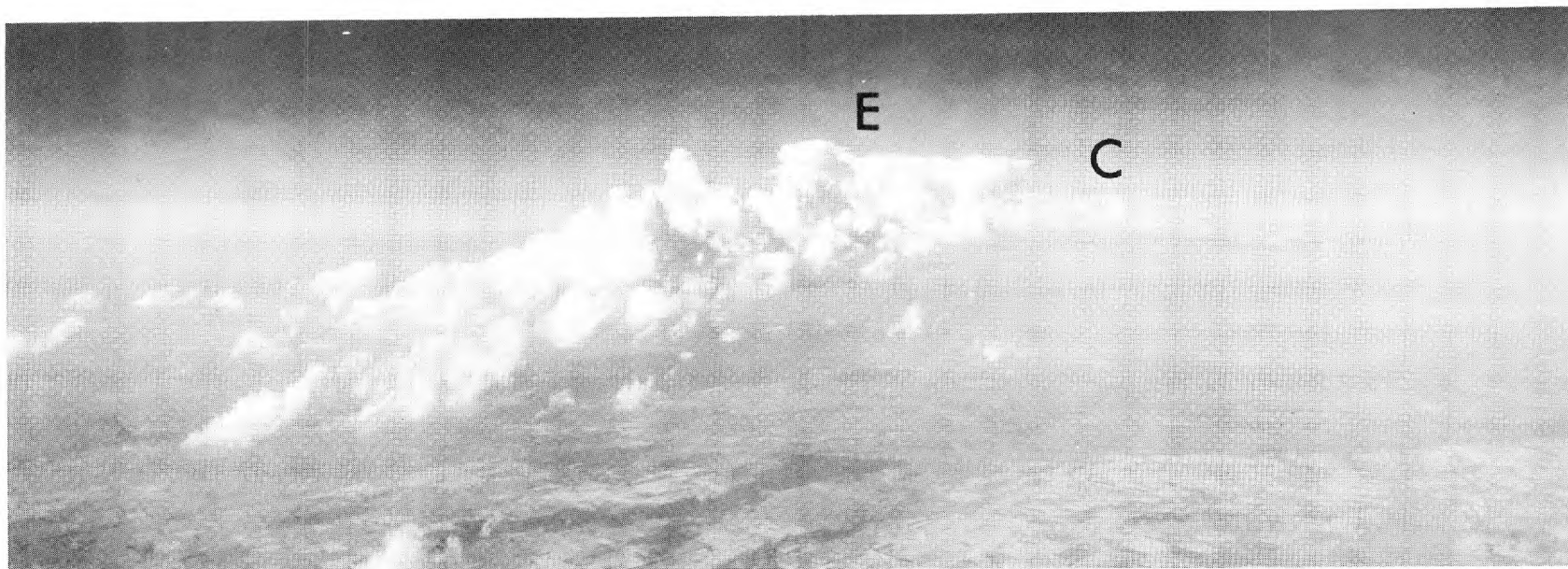


Fig.85. View from 25,000 ft looking NE from position 6, Fig.78, 1632 C.S.T., 26 May 1962. The clouds in the centre (E) are the congestus group seen in the centre of the previous picture, now beginning to produce anvils and partially obscuring the more distant cumulonimbus of group C.

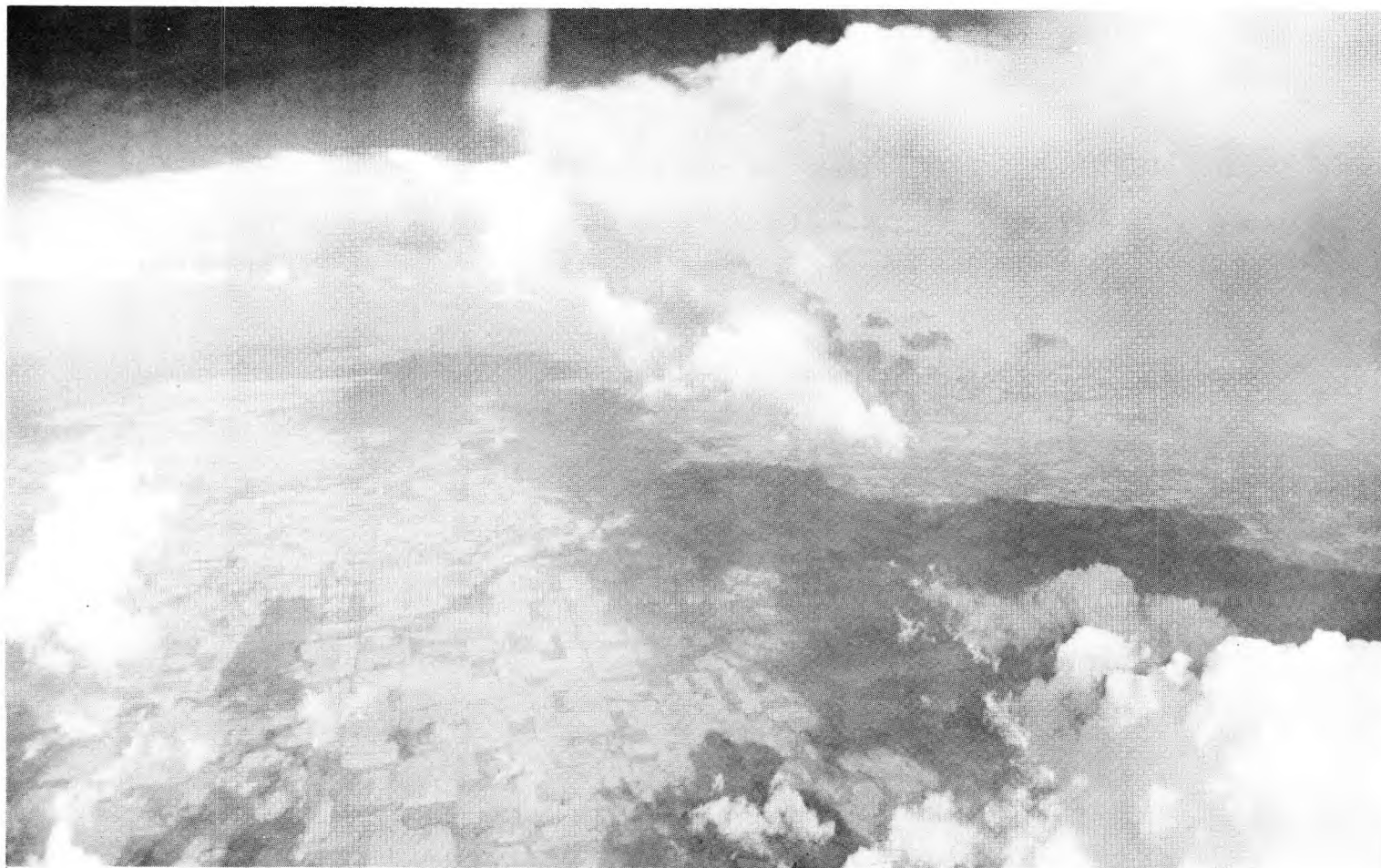


Fig. 86. View from 25,000 ft looking E along dry line from position 7, Fig. 78, 1654 C.S.T., 26 May 1962. On the right is the anvil from the cumulonimbus of group E, previously the congestus shown in the centre of Fig. 85. The group of congestus on the far left lie close to the flanks of group C.

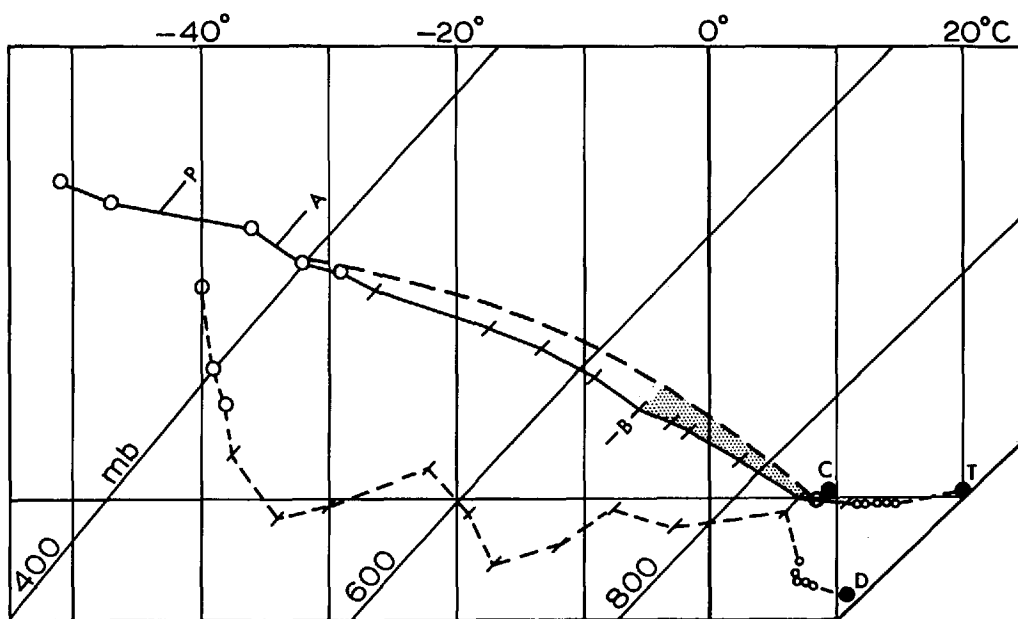


Fig.87. Aircraft sounding inland over England (near Dunstable), 1330 G.M.T., 2 July 1956. The points T, D and C represent respectively the temperature, dew-point and adiabatic condensation level of the screen-level air. The points marked as small circles were obtained from the observations of a nearby tethered balloon at 1500 G.M.T. The cumulus layer is shown by stippling against the saturated adiabatic through the point corresponding to the observed cloud base level, extending up to the level B. The anvil tops of nearby cumulonimbus were at the level A, and the peaks reached level P.

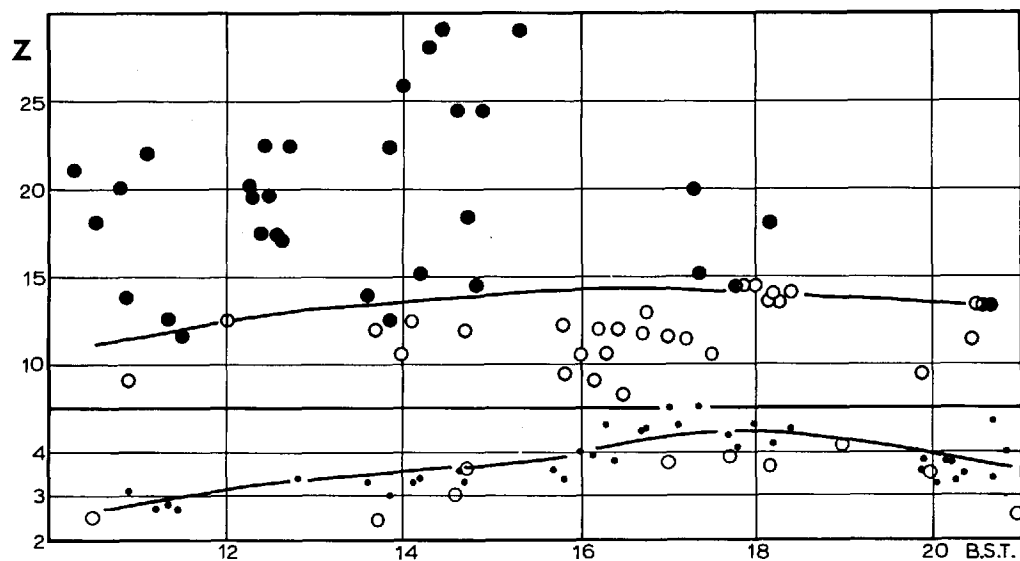


Fig.88. Variation during day (2 July 1956) of height Z (in thousands of feet) of cumulus base (lower part of diagram: dots represent estimates from double-theodolite observations and circles the condensation-level of screen-level air) and of peak tops (upper part of diagram: measurements by double-theodolite and radar of peak heights attained by clouds which fibrillated or produced precipitation shown by full circles. Peak heights of cumulus towers which evaporated without fibrillation of precipitation are shown by open circles).

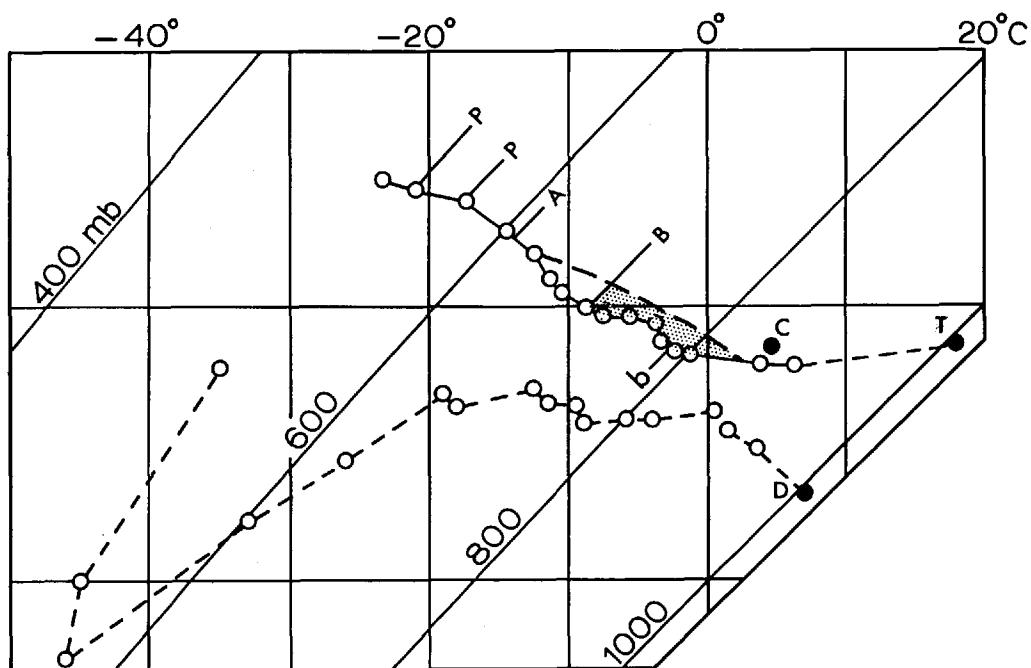


Fig.89. Aircraft sounding inland over England (Bishop's Stortford) 1500 G.M.T., 25 June 1957. The points T, C, D represent respectively the temperature, dew-point and condensation level of the screen-level air. The cumulus layer, shown by stippling against the saturated adiabatic through the point corresponding to the observed cloud base, extends locally up to the level b and more generally to the level B. The anvil tops of small cumulonimbus were observed at the level A, with domes frequently to the level p and occasionally to the level 0.

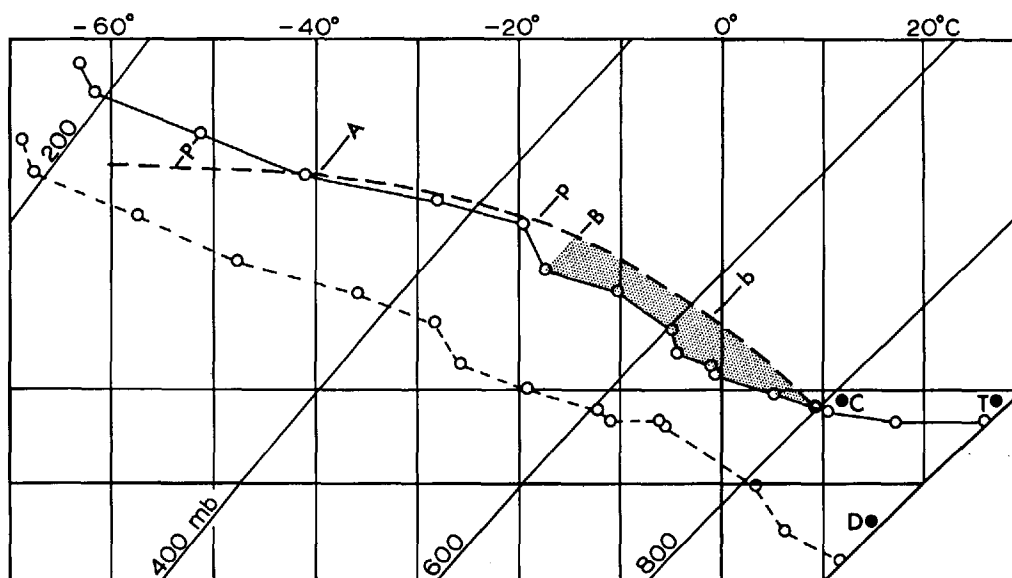


Fig. 90. Radiosonde sounding at Ostersond, central Sweden, 1500 G.M.T., 10 July 1955. The points T, D and C represent respectively the temperature, dew-point and condensation level of air at screen-level. The cumulus layer is shown by stippling against the saturated adiabatic through the point corresponding to the observed cloud base level. The tops of the local cumulus extend up to the level b, and the peak tops over higher ground to the level p. There cumulonimbus were observed with anvil tops at the level A and peak tops to the level P.

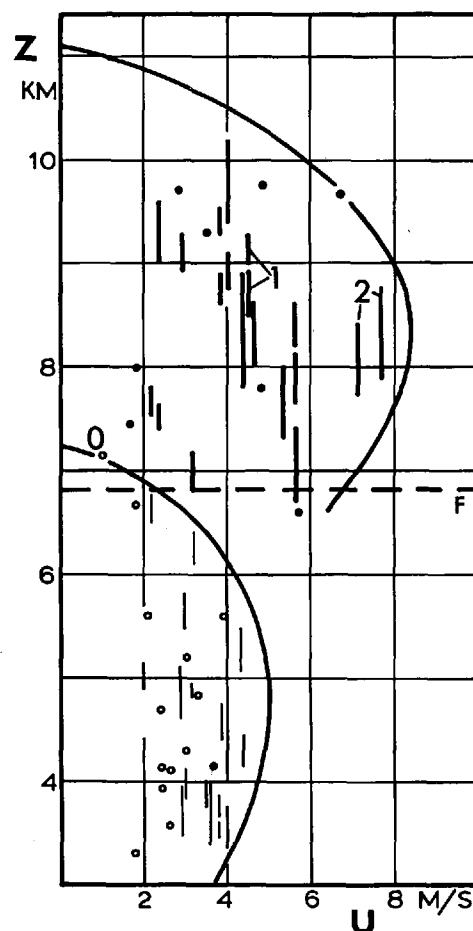


Fig. 91. Observations on 10 July 1955 of rate of rise of emergent cloud tower tops, measured by the double-theodolite method (black dots for precipitating cloud, open circles for cumulus cloud) or from time-lapse films (vertical lines representing mean upward speed over the height interval indicated), all as a function of height Z. The two enveloping curves separate characteristic regimes associated with cumulus and with cumulonimbus. The observations marked 1 represent the first tall cumulonimbus towers produced following the fibrillation of a congestus group represented by the point 0, and the observations marked 2 are of succeeding towers in the same cumulonimbus.

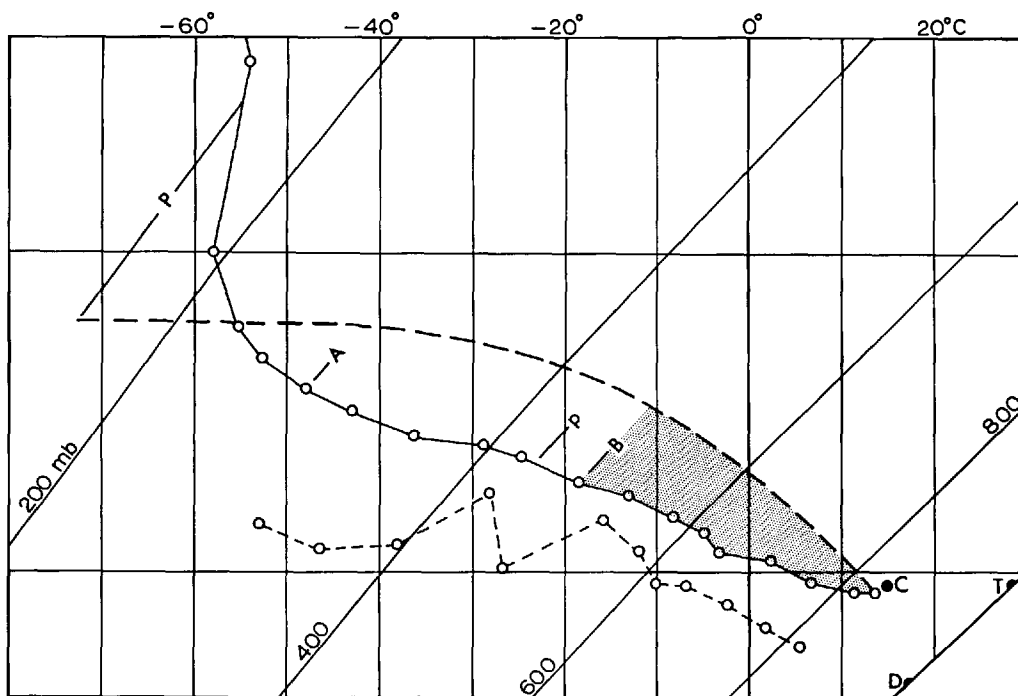


Fig.92. Radiosonde sounding at Crawley (03774) in southern England on 18 June 1957. The points T, D, C represent respectively the temperature, dew-point and condensation level of screen-level air. The stippling against the saturated adiabatic through the point corresponding to the observed cloud base shows the cumulus layer extending up to the level B. Showers developed in the cumulus whose tops attained the level p (at 6700 m), and the cumulonimbus anvils produced had tops at the level A (at 10,000 m), while the highest domes reached the level P (at 13,200 m).

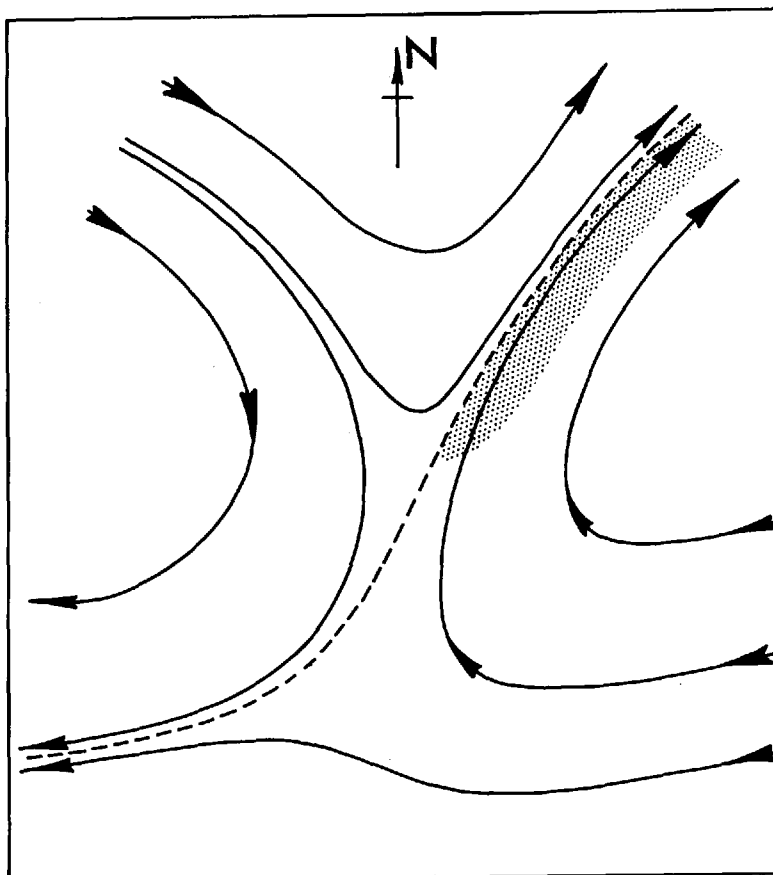


Fig.93. Structure of flow in a large-scale trough, as represented on a relative-flow isentropic analysis. The stippling marks a cloud system where strictly the flow leaves the isentropic surface. The height of the surface is a maximum in the north, so that west of the front (pecked line) the motion is generally descending, and east of it ascending.